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GEOLOGY OF THE EASTERN NAZARETH MOUNTAINS, PALESTINE

I. CRETACEOUS STRATIGRAPHY

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ABSTRACT

In the Eastern Nazareth Mountains in Lower Galilee, Palestine, the Cretaceous is represented by a continuous series of marine deposits ranging from the Upper Albian to the Maestrichtian and possibly to the Danian. The uncertainty of the stratigraphic position of the Vraconian reflects the need of redefining the boundary between the Lower and Upper Cretaceous. The occurrence of "pseudo" flat-pebble conglomerates and the origin of the chert of the Cenomanian deposits are discussed. The mode of recurrent rudistite bioherms in the Lower Turonian is demonstrated. Ammonites characteristic of the Austin chalk of Texas occur in the Coniacian-Santonian beds. Differences between the Maestrichtian of the area studied and of the Palestinian standard section are discussed.

The Cretaceous deposits represent a record of the widespread Upper Cretaceous transgression beginning in late Albian time. Fluctuating conditions in Cenomanian, Turonian, and Senonian time are reflected by the character of both the sediments and faunas.

INTRODUCTION

LOCATION AND TOPOGRAPHY

The area under discussion is situated in the southern part of Lower Galilee, Palestine. The backbone of the Eastern Nazareth Mountains is formed by the Jabal Sih Range, stretching from Nazareth northeastward toward Sajara. Toward the north a gentle-sloping hilly area is bordered by the small Tur'an plain (Sahel Tur'an-Wadi Rumane). To the east the mountain complex merges morphologically into the Galilean Basin with the exception of the

isolated Lubya hill to the northeast. To the south a front range with a steep fault scarp borders the Tabor plain between Jabal Aqaba and Jaba Sartaba. Both the Jabal Sih and Jabal Sartaba merge southeastward into a low-lying hilly area which separates them from the isolated Mount Tabor horst. Westward, toward the Western Nazareth Mountains, an arbitrary boundary line was drawn running from the Wadi Gharbi to Nazareth and from there along the Nazareth-Tiberias highway. The morphology of the Eastern Nazareth Mountains in their major features is controlled by structure.

Field work was carried on from November, 1935, to May, 1936. As a result of political disturbances the final mapping of the southeast part of Mount Tabor could not be finished.

GENERAL STATEMENT

Cretaceous deposits are widely exposed in the area studied, forming a major part of the mountain formations. The areal distribution is less extensive, however, than previously reported.[1] The Cretaceous is represented by a sequence of continuous marine deposits, ranging from the uppermost Lower Cretaceous to the higher Upper Cretaceous. The following divisions of the Cretaceous were recognized: Upper Albian–Vraconian, Cenomanian, Turonian and the Coniacian-Santonian, Campanian, and Maestrichtian of the Senonian. The occurrence of a transitional series between the Maestrichtian and the Lower Eocene beds suggests the presence of Danian equivalents. Supporting faunal evidence, however, is wanting.

UPPER ALBIAN–VRACONIAN

AREAL DISTRIBUTION

The oldest rocks in the Eastern Nazareth Mountains occur in two limited areas at the northern and eastern mountain rim, bordering the Tabor plain. Here the outcrops are located at the base of the fringing fault escarpments where the displacement exposes the deep-

[1] M. Blanckenhorn, "Kurzer Abriss der Geologie Palästinas," *Zeitschr. d. deutsch. Paläst. Ver.*, Vol. XXVI (1912), map; F. J. Fohs, "Geology and the Petroleum and Natural Gas Possibilities of Palestine etc.," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XI (1927), map; P. Range, "Wissenschaftliche Ergebnisse einer geologischen Forschungsreise nach Palästina im Frühjahr 1928," *Zeitschr. d. deutsch. Paläst. Ver.*, Vol. LV (1932), map; L. Dubertret, "La Carte géologique au millionième de la Syrie et du Liban," *Rev. de géog. phys. et de géol. dynam.* (Paris, 1933), map.

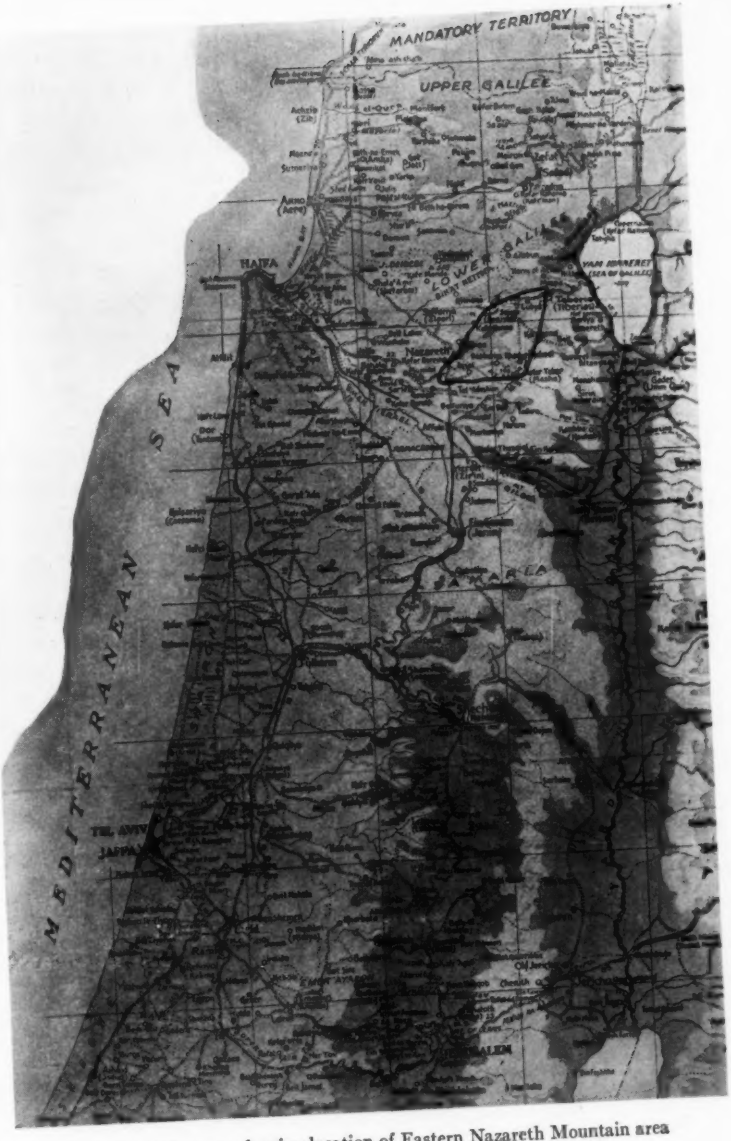

FIG. 1.—Index map showing location of Eastern Nazareth Mountain area

est part of the dissected crest of the "Dabburiya anticline." The larger of the exposures lies east of the village of Iksal. The beds occupy the basal section of the mountain escarpment on either side of the lower course of the Wadi Sartaba. The smaller outcrop is located at the northwestern base of Mount Tabor, where the rocks are exposed in the road cut above the southern edge of the village of Dabburiya.

The outcrop near Iksal exposes a greater vertical range of the Vraconian than that at Mount Tabor. In both sections, however, only partial profiles can be observed because of the extensive masking of the rocks by secondary calcareous incrustations, the so-called "nari," and by soils.

PREVIOUS WORK

The outcrop northeast of Iksal was reported previously by G. S. Blake in 1928 and 1936. In 1928[2] the deposits were referred to the Lower Cretaceous. Subsequently[3] the age was established as Upper Albian (Vraconian).

LITHOLOGY

The Vraconian rocks comprise limestones, shaly limestones, and clay beds, all of which are ferruginous. The color of the rocks is commonly deep brown, occasionally buff to green. The clay content appears to decrease toward the top of the formation, as far as can be ascertained from the partial sections. Iron, the dominant coloring agent of the rocks, is commonly concentrated in the form of limonite, as fillings of rock cavities and worm borings and as replacement of mollusk shells. Glauconite grains up to $\frac{1}{2}$ mm. in diameter are common in the fossiliferous shaly and coarse limestone beds.

THICKNESS

In the Wadi Sartaba the thickness of the exposed rock of the Upper Albian and Vraconian can be estimated as 240 feet.

FAUNA

Fossils occur in both of the outcrops. Ostreidae and Pectinidae are common at the base of Mount Tabor.[4] In the Wadi Sartaba

[2] *Geology and Water Resources of Palestine* (Jerusalem, 1928), p. 11.

[3] *The Stratigraphy of Palestine and Its Building Stones* (Jerusalem, 1936), p. 62.

[4] No specific identification could be made, since the collected specimens were among those lost in the Bedouin camp of Arab Sbeih.

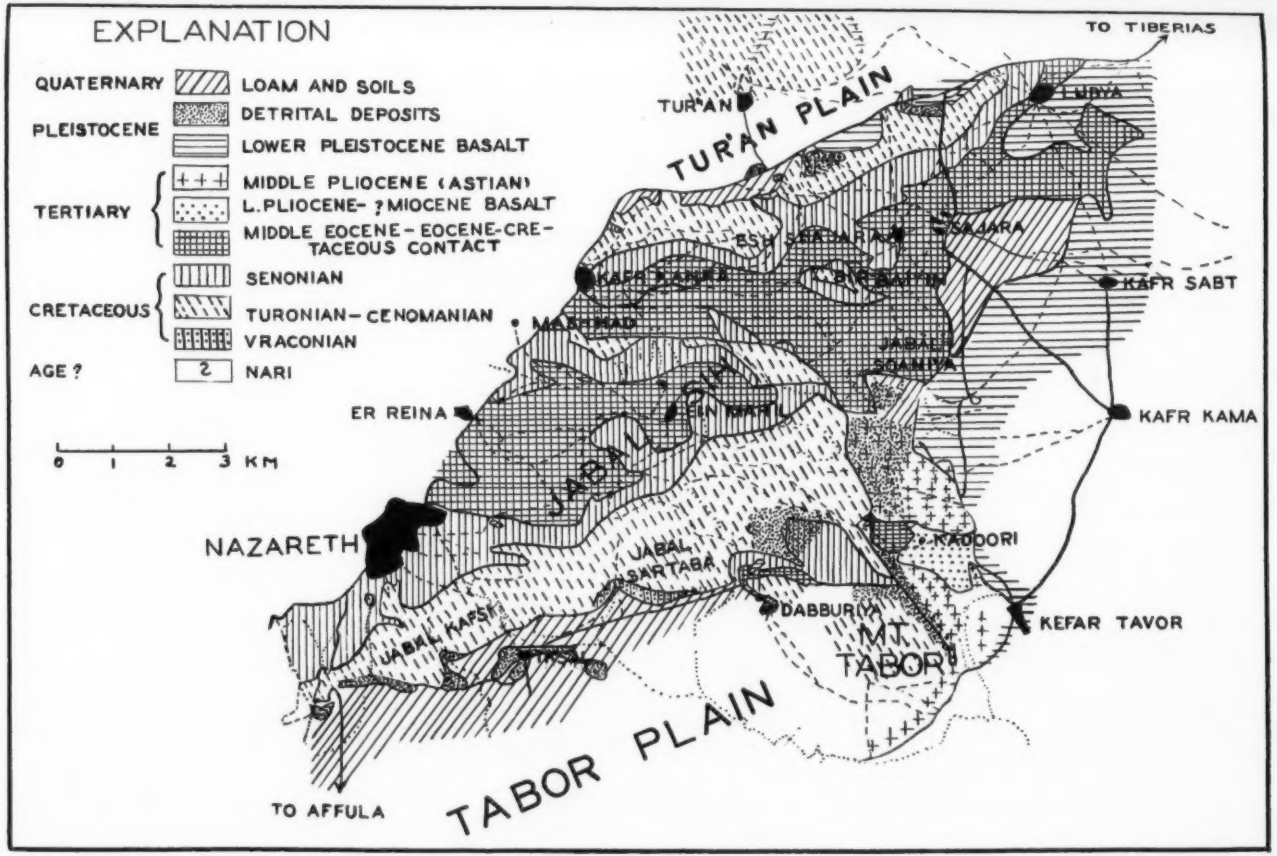

FIG. 2.—Geological map of the Eastern Nazareth Mountain area

section several fossiliferous beds occur. The highest of the fossil zones observed is crowded with *Orbitolina conica;* and in a zone 6.5 meters below, *Exogrya flabellata* outnumbers all other faunal elements. Blake[5] reported *Pholadomya depacta* from the Wadi Sartaba section. To this can be added the following forms:

Orbitolina conica d'Archiac	*Exogyra flabellata* Goldfuss
Echinoid remains	*Tylostoma* cf. *induratum* Conrad
Orbiculoidea sp.	*Cerithium* sp.
Anisocardia orientalis Conrad	*Pseudomelania* sp.
Lima (Radula) *naamensis* Whitfield	Fish teeth (conchivorous)

The specifically identified forms are all known to occur in the Upper Albian–Vraconian of Syria and partly in Palestine.

CORRELATION

Anisocardia orientalis and *L. naamensis* are restricted to the Upper Albian–Vraconian. *Pholadomya depacta* (found by Blake) and *T. induratum* extend in both Palestine and Syria down to the Lower Albian, and *O. conica* ranges up into the Lower Cenomanian and E. *flabellata* as far as lower Senonian, becoming cosmopolitan during Cenomanian time. The transitional character of the faunal assemblage agrees well with the original definition of the Vraconian. Renevier[6] defined the Vraconian as the transitional zone between the Albian and the Cenomanian (Lower-Upper Cretaceous contact). Subsequently opinions regarding its stratigraphic position have been divided. Some students of the Cretaceous included the Vraconian in the Lower Cenomanian;[7] others considered it as Upper Albian.[8] The same uncertainty exists concerning the position of the Vraconian in the eastern Mediterranean area, as can be seen from the correlation of the Lebanon deposits. H. Douvillé and G. Zumoffen[9] considered the Vraconian of the Mediterranean as the equivalent of the Upper

[5] *The Stratigraphy of Palestine . . . .*, p. 64.

[6] Cited by A. de Lapparent, *Traité de géologie* (5th ed.; Paris: Masson et Cie, 1906), Part III, pp. 1297, 1353.

[7] *Ibid.*

[8] A. Geikie, *Text Book of Geology* (New York: P. F. Collier & Son, 1902), Part IV, pp. 1560, 1564.

[9] "Etudes sur les rudistes: rudistes de Sicile, d'Algérie, d'Egypte, du Liban et de la Perse," *Mém. soc. géol. France*, Vol. XVIII (1910).

Albian on the basis of faunal evidence. Blanckenhorn[10] regarded it first as the base of the Cenomanian but later included it in the lower Cretaceous. In recent studies of the Syrian Cretaceous, Dubertret[11] accepted Douvillé's definition; and Blake,[12] who discovered Vraconian deposits in Palestine, followed the same correlation. The usage of headings like "Albian" (Vraconian)[13] demonstrates clearly the need for clarification of the stratigraphic problems involved. The present writer is well aware of the fact that the small faunal assemblage, as well as the lack of the characteristic zone fossils—for example, *Knemiceras syriaca* and *Heteraster delgadoi*—does not permit the generalization of apparent transitional character as deduced from the few faunal elements. The question whether the Vraconian stage should be retained as a stratigraphic unit must be answered by a restudy of the European type sections. With respect to the Syrian-Palestine deposits it appears likely that a transitional fauna of the character originally defined by Renevier may well exist, since the Upper Albian represents the initial phase of a continuously spreading sea transgression culminating in the Upper Cretaceous. Studies of fossiliferous Upper Albian–Lower Cenomanian sections in Syria and Palestine are greatly desired both for the determination of the vertical range of all the faunal elements and for the succession of faunal assemblages. The results of such a study would demonstrate whether or not a transitional Vraconian stage is recognizable, which should be separated from the Upper Albian. Until such a study has been carried out, the deposits under discussion are provisionally classified as Upper Albian and Vraconian.

CENOMANIAN

GENERAL STATEMENT

The Upper Albian–Vraconian strata are conformably overlain by a great thickness of Cenomanian rocks. The outcrops are commonly characterized by steep escarpments and narrow wadi gorges, resulting from the hardness of the predominant dolomitic rock.

[10] "Die fossilen Gastropoden und Scaphopoden der Kreide von Syrien-Palästina," *Palaeontographica*, Vol. LXIX (1927), p. 103; "Die Bivalven der Kreideformation von Syrien-Palästina," *ibid.*, Vol. LXXXI, Part A (1934), p. 164.

[11] *Op. cit.* [12] *The Stratigraphy of Palestine . . . .*, p. 64. [13] *Ibid.*

AREAL DISTRIBUTION

Continuous belts of exposures occur at the mountain flanks bordering the Tabor and Tur'an plains. The southern mountain escarpment facing the Tabor plain is composed of Cenomanian rocks, which extend eastward into the hilly area between Mount Tabor and Jabal Sih. Toward the Galilean Basin this outcropping belt is delimited by the Tabor fault. To the south the Mount Tabor horst is entirely built up of Cenomanian rocks, which at the northwestern flank are separated from the main mass by the Arab Sbeih fault. Extensive exposures border the Tur'an plain. The fault scarps of the Jabal Tur'an to the north and the low-lying belt at the mountain flank to the south consist of Cenomanian strata. Scattered outcrops occur in the mountain interior along the northern fault scarp of the Jabal Sih and in the eroded core of the domelike anticline of Bir Baiyin. In the hilly area north of Mount Tabor two small exposures are located in the cuts of the Wadi M'qata.

PREVIOUS WORK

Reviewing previous correlations of the sequence considered as Cenomanian, O. Fraas[14] reported that the Cretaceous rocks on the top of Mount Tabor are comparable with the Maestrichtian limestones. On the first detailed map of Palestine, Blanckenhorn[15] referred the strata at the northern and southern mountain rim to the Cenomanian-Turonian, both of which were mapped as one unit. In a subsequent report Blanckenhorn[16] recorded the presence of both the Cenomanian and Turonian at Mount Tabor. Since the youngest deposits of the Mount Tabor section are late Cenomanian in age, it is evident that Blanckenhorn referred parts of the Upper Cenomanian deposits to the Turonian. On Range's[17] geological sketch map of Galilee the age of the same areas marked by Blanckenhorn as Cenomanian-Turonian was limited to Turonian. This change was based

[14] *Aus dem Orient*, I. *Geologische Beobachtungen am Nil, auf der Sinai-Halbinsel und in Syrien* (Stuttgart: Ebner & Seubert, 1867), p. 71.

[15] "Kurzer Abriss der Geologie Palästinas," *op. cit.*, map.

[16] *Syrien, Arabien und Mesopotamien, Handbuch der reg. Geol.*, Vol. V, No. 17 (1914), p. 192.

[17] *Op. cit.*

on L. Picard's[18] age determination of the western part of the southern belt. Picard's suggestion of a Turonian age was based on lithologic criteria and the contact relation to the overlying rocks, which he regarded as Santonian in age (see discussion of the Turonian). Detailed Cenomanian sections from the Jabal Aqaba and Mount Tabor were recorded by Blake.[19] The strata below the so-called "Mizi helu"[20] were referred to the Cenomanian (Jabal Aqaba), and at Mount Tabor the sequence was subdivided into an Upper and Lower Cenomanian complex. Rudistites, reported from the basal "Upper Cenomanian" of the latter section, are the first recorded Cenomanian fossils from the entire Nazareth Mountains.

LITHOLOGY

The Cenomanian deposits comprise a well-bedded series of brown to buff-colored dolomites, dolomitic limestones, and limestones. The dolomites and dolomitic limestones form the dominant rock component. Pure limestones occur irregularly interbedded between the dolomites. Lateral gradation of dolomites into limestones and vice versa are common phenomena. The variability of the different rock components vertically is indicated by comparison of three detailed sections, two given by Blake[21] from the Mount Tabor and the Jabal Aqaba and one given by Picard[22] from the Western Nazareth Mountains, northeast of Ginegar.

Locally, slightly argillaceous limestones, thin shale bands, and marls are present. The shale bands occur in the "presumably upper part of the Cenomanian," in a bore section[23] in the southern Tur'an plain. The occurence of marls was reported from the basal Upper Cenomanian of Mount Tabor by Blake.[24] The argillaceous deposits are not limited to a single zone but occur locally in various parts of the sequence.

[18] "Zur Geologie der Kischon-Ebene," *Zeitschr. d. deutsch. Paläst. Ver.*, Vol. LI (1928), pp. 7–8.

[19] *The Stratigraphy of Palestine* . . . . , p. 58.

[20] See p. 834.

[21] *The Stratigraphy of Palestine* . . . . , p. 58. [22] *Op. cit.*, p. 8.

[23] Bore log and stratigraphic interpretation, communicated by Blake in a letter May 7, 1937.

[24] *The Stratigraphy of Palestine* . . . . , p. 58.

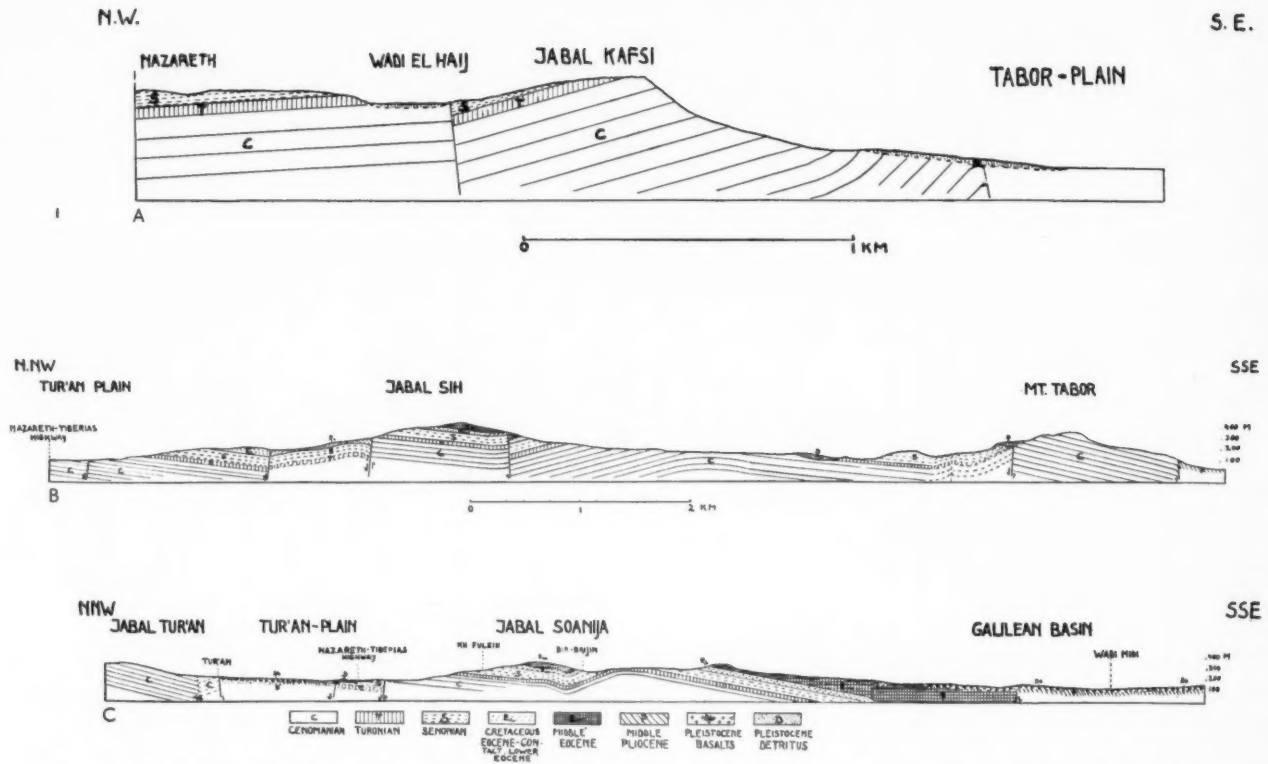

FIG. 3.—*A*, geologic cross section from the town of Nazareth to the Tabor plain; *B*, geologic cross section from the western part of the Tur'an plain to the southeast foot of Mount Tabor; *C*, geologic cross section from Jabal Tur'an to the Wadi el Midi (Galilean Basin). In all sections vertical and horizontal scales are the same.

Intraformational conglomerates and breccias were found locally in the upper part of the Cenomanian sequence, in the highway cut south of Nazareth, in the Wadi Sartaba, south of Ayun Shain, at the southeastern flank of Mount Tabor, and in Cenomanian boulders in the Tur'an plain. The intraformational conglomerates of Mount Tabor and in the Tur'an plain appear noteworthy for the derivation of the pebbles. At Mount Tabor the pebbles occur in a zone which is crowded with silicified pelecypod and gastropod shells. A large number of the pelecypod shells are fragmentary, well rounded, and abraded to such an extent that the original surface markings have been nearly obliterated. The associated pebbles are flat and show rounded edges like those of flat-pebble conglomerates. This association with the abraded shell fragments suggests that the pebbles may have been originally shell fragments whose surface markings have been entirely removed. No shell structure was found in sectioned pebbles because of the obliteration of structural details.

Loose boulders in which the pebbles occur are apparently derived from the eastern flank of the Jabal Tur'an, but a brief search in this region, which lies outside of the area studied, failed to locate the deposits *in situ*. Most of the larger constituents of the boulders are flat pebbles and fragments of colonial corals of the Helioporacea type. The interstitial matrix is predominantly composed of small pebbles, grading into a "pebble sand." There are all gradations from completely polished flat pebbles to similar-shaped fragments of uncorroded coral colonies. Evidence that the flat pebbles were derived from corallite fragments was found in the presence of coral structure in small depressions on the otherwise smooth and polished pebbles and by the presence of coral structure in sectioned specimens entirely free of such surface indications. The shaping of the pebbles was favored by the longitudinal cleavage along separating walls of tube-shaped individual corallites. It is proposed to call this pebble type "pseudo" flat pebble.

Pseudo flat pebbles may be derived from fragments of colonial corals of the morphological[25] type described, as well as from pelecy-

[25] On the basis of the closely related morphology similar pseudo flat-pebble conglomerates should be found around the Paleozoic "bioherms" of *Favosites* and related coral forms and possibly of *Chaetetes*.

pods with minor curvature, by abrasion and corrosion in a rough-water environment. The assemblage of densely crowded coral fragments and pseudo flat pebbles in the boulders suggests derivation from the talus material of a coral reef. If so, the source area (the reef) and the deposition area of the pseudo flat pebbles (the adjacent sloping edge of the reef) were not far from one another. The resem-

FIG. 4.—Jabal Kafsi with Wadi Kafsi

blance to the typical flat pebble was achieved by abrasion[26] in the rolling sea which entirely obliterated the surface markings of the

[26] The extent of the abrasion period, necessary to obliterate the surface markings of the coral structure, appears to depend largely upon the part of the reef from which the particles have been derived. Fragments derived from the living zone of the reef, where the relatively strongest morphological differentiation exists, have to undergo the longest corrosion process. Alterations take place in the dead part of the reef, partly in the form of solution and erosion and partly in the form of cementation and incrustation. These processes result in a flattening of the surface of the dead reef. The abrasion period of fragments derived from the dead part of the reef will be considerably shorter, especially if the particles are relatively free of secondary borings common in this part of the reef.

coral fragments. The association of the pseudo flat pebbles with small well-rounded and polished pebbles suggest a relatively long phase of abrasion. The mode of forming of the pseudo flat pebbles explains, however, the unusual assemblage of flat and well-rounded pebbles.

Chert occurs in separate areas at different horizons. At the southern fault scarp between Jabal Aqaba and Jabal Chamsa the cherts are confined to the lower part of the Cenomanian sequence. There is one continuous zone which consists of several bands of nodular and lenticular cherts intercalated in the dolomite. To the west the zone ends above in a thin jasper bed (Jabal Kafsi). At the western margin of the occurence, the local development of a second zone of nodular chert was observed, 10 meters higher up. At Mount Tabor and in the hilly area to the north at the Wadi Maqam esh S'hab cherts occur in the upper part of the Cenomanian sequence. Several bands of nodules and large flat lenticles (Wadi Maqam esh S'hab) are developed.

Generally, the nodules appear to be devoid of organic inclusions, although at the occurence between Jabal Aqaba and Jabal Chamsa abundant fossiliferous nodules were found. The majority of these contain sponge spicules and a few foraminifera. Radiolarians are also common. The abundance of the spicules in the different nodules varies, some being crowded with spicules which form layer-like masses up to 19 mm. in thickness. There is no lateral, marginal decrease in the thickness of the spicule masses; one gains the impression of a laterally interrupted layer. The spicule masses show lack of orientation. Monactinellid elements among the spicules are dominant. Corrosion marks on radiolarian skeletons and spicules with enlarged canals are common.

A few nodules show a meshwork-like structure. In some places this meshwork appears to be formed by unoriented sponge spicules, whose outlines are indistinct. Locally, a few completely preserved spicules were detected.

An associated chert type is characterized by concentrically banded nodules. They are commonly devoid of organic remains. In one case, however, two minute gastropods were found in the center be-

FIG. 5.—Southern fault scarp of Jabal Kafsi, as seen from the Tabor Plain

tween indeterminable fragments of megascopically visible fossils and molds of echinoid remains on the surface.

The fossil content of the chert nodules furnishes some information about the depth of water and sedimentation conditions and, as far as the sponge remains and radiolarians are concerned, about the agent precipitating the silica. The abundance of sponge spicules suggests bioherms of siliceous sponges. Crowding of the sponges seems to be indicated by the presence of the few minute gastropods and echinoids. "Dwarf" faunas commonly are associated with such types of sponge bioherms and have been explained as adaptations to the small cavities and passages of the bioherms.[27] The predominance of monactinellid elements among the spicules permits the conclusion that the reefs developed in comparatively shallow water. The lack of orientation of the spicule masses suggests a still-water environment. The rate of sedimentation must have been slower than the rate of disintegration of the soft parts of the sponges, since detached spicules alone are present. Spicules with enlarged axial canals and solution marks on radiolarian skeletons indicate a sufficiently long post-mortem exposure to permit incipient solution by the chemical agents of the sea water which act on amorphous silica. The nodules exhibiting a meshwork-like structure seem to represent the fixation of an advanced stage of solution. If so, this type can be regarded as a transitional stage toward complete solution of the exposed skeleton parts. The variable degree of solution indicates that it depends upon certain local conditions, which vary from place to place. Possibly, the degree of surface exposure of the skeleton material in various parts of the deposit may be the controlling factor.

G. M. Lees[28] in 1928 said, in a general discussion of the chert beds of Palestine, that "the silica appears to have been precipitated directly from the sea water and not through the agency of silica secreting organisms, of which there is a notable lack." But the abundance of sponge spicules and the presence of radiolarians indicate that the silica of the chert occurrence under discussion was precipitated by

[27] E. Dacqué, *Grundlagen und Methoden der Palaeogeographie* (Jena: Gustav Fischer, 1915), p. 236.

[28] "The Chert Beds of Palestine," *Proc. Geol. Assoc. London*, Vol. XXIX (1928), p. 456.

silica-secreting organisms. It may be objected that the quantity of observed sponge spicules and radiolarians is not sufficient to account for all the silica of the chert nodules. The quantitative deficiency expressed by the lack of organic remains in parts of the chert zone must not be regarded as counterevidence. L. Cayeux[29] cites a number of recent observations which indicate that sponge spicules can be easily dissolved on the sea floor if long enough exposed. Evidence of much solution has already been presented. The nodule type, containing layer-like masses of sponge spicules, indicates that originally continuous layers of spicules were present. The dolomites and dolomitic limestones in which the nodules are now enclosed are free from siliceous material. It is therefore evident that a considerable portion of the original spicule layers has been dissolved, the nodules being local remnants of the former spicule layers which were cemented and enveloped by silica. The latter was derived from the desilicified parts of the former skeleton layers by local solution and lateral migration. The limited extent of the chert horizons both horizontally and vertically could correspond with the original extent of the bioherms.

Whether this chertification took place entirely on the sea floor or epigenetically cannot be decided on the basis of the collected data. The concentrically banded cherts indicate considerable duration of the forming process. This chert type has been explained by A. Schwartz[30] and N. L. Taliaferro[31] as representing a separate generation of chert for each of the bands which are not diffusion rings ("Liesegang rings"), as thought by W. Winkler.[32] How rapidly the formation of these bands followed each other remains uncertain. It is also doubtful if the forming of the nucleus of the banded chert was contemporaneous with the forming of the spicule-bearing nodules. Since the few fossils found had a calcareous skeleton, the banded

[29] "Les Roches sédimentaires de France: roche silicieuses: la carte géologique détaillée de la France," *Mém. Paris* (1928), p. 565.

[30] "Zur Bildung der Feuersteine," *Senckenbergiana*, Vol. XI (1929), pp. 56–57, 159–92.

[31] "Contraction Phenomena in Cherts," *Bull. Geol. Soc. Amer.*, Vol. XLV (1934), pp. 189–232.

[32] "Über die Bildung mesozoischer Hornsteine: Ein Beitrag zur Sedimentpetrographie der Julischen Alpen," *Tscherm. min. u. petrog. Mitt.*, Vol. XXXVIII (1925).

chert could be supposed to have formed after the desilicification of the sponge-spicule layers.

The fossiliferous bed at the southeastern flank of Mount Tabor and the loose boulders in the Tur'an plain mentioned in connection

FIG. 6.—Mount Tabor seen from the east slope of Jabal Sartaba

with the description of the pseudo flat-pebble conglomerate consist of silicified limestones. They correspond to the "secondary quartzites" of Lees.[33]

THICKNESS

The total thickness of the Cenomanian in the Wadi Sartaba section is 450 meters.

FAUNA

The Cenomanian deposits are sparingly fossiliferous. Fossils occur mainly in form of casts and molds and are generally poorly preserved. Lenslike concentrations of fossil remains were found sporadically.

The earliest remains observed in these Cenomanian strata occur

[33] *Op. cit.*, p. 455.

in chert nodules at the southern mountain escarpment. The faunal assemblage representing a sponge bioherm has been mentioned above in the discussion on the origin of the chert nodules. The radiolarians form the only faunal element which may permit specific identification. They have not as yet been studied.

A prolific fauna was found in the highway cuts (highway: Nazareth-Affula) on the southern mountain escarpment. The fossils occur in a lenslike body which extends horizontally for nearly 40 meters and has an exposed thickness up to 1 meter. The fossil deposit occurs 190 meters below the top of the Cenomanian and is stratigraphically higher than the fossiliferous nodules mentioned above. The fauna contains the following forms:

? Trochosmilia sp.
Serpula sp. nov.
Avicula cenomanensis d'Orbigny
Nucula sp.
Leda sp.
Cucullaea sp.
Arca sp. nov.
Cardita nicaisea Coquand
Astarte sp.
Cardium sp.
Protocardia cf. *coquandi* Seguenza
Anisocardia aquilina Coquand
Dosinia delettrei Coquand
Venus reynesi Coquand
Dentalium sp.
Delphinula porteri Blanckenhorn
Natica sp.
Turritella sp.
Nerinea cf. *abeihensis* Blanckenhorn
N. cochleaeformis Conrad
N. (*Nerinella*) cf. *schicki* Fraas
N. longissima Reuss
Cerithium cf. *intercalatum* J. Boehm
Cerithium aff. *sejunctum* Zekeli
Mesostoma shalemi Blanckenhorn
Rostellaria cf. *bekariensis* Blanckenhorn
Hastula(?) *aciformis* Blanckenhorn
Actaeonella salomonis Fraas
A. syriaca Conrad[34]

The bulk of the specifically identified forms is confined to the Cenomanian, including the following: *A. cenomanensis*, *C. nicaisei*, *A. aquilina*, *D. porteri*, *N. longissima*, *M. shalemi*, *H.*(?) *aciformis*, and *A. syriaca*. Of these forms, H.(?) *aciformis*, *D. porteri*, and *A. syriaca* occur exclusively in the Lower Cenomanian. *H.*(?) *aciformis*, *N. cochleaeformis*, and *A. syriaca* are forms which Blanckenhorn[35] considers as index fossils of the Lower Cenomanian. According to Blake,[36] however, *H.*(?) *aciformis* and *N. cochleaeformis* are

[34] Blanckenhorn, "Die Bivalven . . . . ," *op. cit.*, summarizing table, used for all correlated age ranges.

[35] "Geologie Palästinas nach heutiger Auffassung," *Zeitschr. d. deutsch. Paläst. Ver.* Vol. LIV (1931), p. 9.

[36] *The Stratigraphy of Palestine* . . . . , p. 54.

"characteristic fossils" of the upper complex of the Lower Cenomanian. *H.*(?) *aciformis* ranges from the Vraconian[37] to the Upper Cenomanian.[38] Consequently, it cannot be considered as an index fossil.

Of the specifically identified forms, 83 per cent have a pronounced Mediterranean character. Of these, 44 per cent are restricted to the eastern Mediterranean countries. There is a subordinate influx of European forms, indicated by the presence of *A. cenomanensis*, *N. longissima*, and a form which is closely allied with *C. sejunctum*.

The remaining fossils, collected from various horizons in the area studied, were found to be either too poorly preserved for specific identifications or noncharacteristic, owing to their considerable vertical range.

Fossils differing from those collected in the Eastern Nazareth Mountains were found in loose, silicified boulders in the wadis of the Tur'an plain. These boulders, which were probably brought down from the eastern flank of Jabal Tur'an, have been mentioned above in the discussions on the pseudo flat-pebble conglomerates and the silicified limestones. Two types of boulder occur, which differ lithologically and in their faunal composition.

The light-gray, dense type yields the following fossils:

Praealveolina sp. (cf. *Praelveolina cretacea tenuis* Reichel)
Sponge spicules (monactinellid) rare
Anatina(?) *orientalis* Whitfield
Patella sp.
Ceritella bilineata Billings
Nerinea dubia Blanckenhorn
N. sp.
Cerithium sp.
Actaeonella sp.

The wide range of the few specifically identified peleypods and gastropods—that is, from Aptian to Cenomanian—does not permit a definite age determination of these fossils. The presence of the genus *Praealveolina*, however, which is unknown from pre-Cenomanian deposits, according to M. Reichel,[39] establishes a Cenomanian age of the fauna.

[37] Blanckenhorn, "Die Bivalven . . . . ," *op. cit.*, summarizing table.

[38] Blake, *The Stratigraphy of Palestine* . . . . , p. 54.

[39] "Etude sur les alvéolines 2," *Mém. Soc. pal. Suisse*, Vol. LIX (1937), pp. 135–36.

The faunal composition of the porous pseudo flat-pebble bearing type is different:

Individual corals
Helioporacea (cf. *Polytremaeis*)
Nerinea sp.
Cerithium sp. (cf. *Pyrazus elias* Boehm)
Actaeonella sp.

Fragments of colonial corals form the bulk of the fossils. Although the fauna of both boulder types is different in general, the presence of a *Nerinea* type occurring in both suggests that they are derived from the same stratigraphic unit. Since the listed *Certhium* sp. is only closely related but not identical with *C. elias*, an index fossil of the Lower Cenomanian, no definite statement is possible as to whether the faunal assemblage is Lower or Upper Cenomanian.

CORRELATION

The separation of the Cenomanian from the underlying and overlying deposits was, of necessity, based on lithologic criteria, since fossils are missing in the boundary zones. In the Wadi Sartaba, the contact with the underlying Upper Albian–Vraconian rocks is observable. The boundary line was arbitrarily drawn there at the base of the dolomitic limestone series, which is typical of the Cenomanian rocks. In some places the boundary between the Cenomanian and Turonian beds is sharply marked. This is true where pink mottled Cenomanian dolomites are overlain by the so-called "Mizi helu" facies, which is Turonian in age. Commonly, however, the dolomites grade upward into crystalline limestones which resemble the Upper Turonian facies. A white marble-like coarse limestone, the so-called "Meleke," [40] occurs locally, at the top of the presumably Upper Cenomanian in the bore section of the Tur'an plain and south of the Khallet et Tayun on the path to Iksal. This facies represents in the Carmel area the uppermost Cenomanian[41] and in the Jerusalem area the basal Lower Turonian[42] on the basis of faunal evidence.

The fauna cited above affords the only means of estimating the

[40] Arabic term meaning the "royal stone."

[41] Blake, *The Stratigraphy of Palestine* . . . . , p. 56; J. Vroman, "Geology of the Region of Southwestern Carmel (Palestine)" (Ph.D. dissertation, University of Utrecht, 1938), p. 7.

[42] Blake, *The Stratigraphy of Palestine* . . . . , p. 51; Picard, "The Geology of New-Jerusalem," *Bull. V., Geol. Dept. Hebr. Univ.* (1938).

approximate proportional thickness of the Lower and Upper Cenomanian series. On the basis of the faunal data presented, the strata in which the fauna occurs must be considered as late Lower Cenomanian in age. The stratigraphic position, 190 meters below the top of the Cenomanian, indicates that the underlying 250 meters of the deposit (at least) are Lower Cenomanian in age. The exact thickness of the Upper Cenomanian remains uncertain. Blake[43] reports a thickness of 550–750 feet for the Upper Cenomanian of northern Palestine, which indicates an average thickness of 220 meters. In the northern part of Mount Carmel, which lies about 20 kilometers west of the region studied, the thickness of the Upper Cenomanian beds does not exceed 100 meters, according to the same report. It is therefore probable that, of the 190 meters of strata overlying the late Lower Cenomanian fauna, most is late Cenomanian in age. The exclusive presence of the previously mentioned intraformational conglomerates and breccias in the upper 160 meters of the sequence appears to represent additional evidence for a correlation of most of these strata with the Upper Cenomanian. They indicate similar shallow-water conditions as generally observable in the Upper Cenomanian deposits of Palestine-Transjordania.

In conclusion, the Cenomanian sequence is lithologically a predominantly dolomitic series with irregularly interbedded limestones. There exist subordinate, local facies variations. These are indicated by the occurrence of chert nodules and argillaceous limestones in both the lower and upper part of the sequence and by intraformational conglomerates and breccias exclusively in the upper part. On the basis of a late Lower Cenomanian fauna and indications of shallow-water conditions a rough separation of the Lower and Upper Cenomanian strata has been attempted. Additional faunal evidence, however, is necessary for an exact stratigraphic analysis.

TURONIAN

AREAL DISTRIBUTION

The Turonian formation, in contrast to the underlying Cenomanian, is comparatively thin. It crops out generally as a narrow band of well-bedded strata which conformably overlie the Cenomanian

[43] *The Stratigraphy of Palestine . . . .*, pp. 54, 58–59.

deposits in all the cited Cenomanian exposures, with the exception of the Mount Tabor horst and the two small outcrops in the cuts of the Wadi M'gata. The Turonian was mapped with the Cenomanian as one unit.

FIG. 7.—Southern escarpment of the Eastern Nazareth Mountains, bordering the Tabor Plain. From the northwest slope of Mount Tabor.

PREVIOUS WORK

In the area studied, Turonian beds were first separated by Blake,[44] who distinguished in a section from the Jabal Sih: "Mizi helu: Upper Turonian, ?Meleke with *Durania*: Lower Turonian." The occurrence of *Nerinea requieniana* d'Orbigny in the "Mizi helu [Turonian]" of the Jabal Aqaba is cited in the same report.[45]

SUBDIVISIONS

In the following the Turonian sequence is divided into a lower and upper complex. The subdivisions are based on lithologic and faunal criteria.

[44] *Ibid.*, p. 39. [45] *Ibid.*, p. 58.

A. LOWER COMPLEX

Lithology.—The predominant rock type of the well-stratified lower complex is the so-called "Mizi helu,"[46] a dense brown- and pink-mottled limestone, cream-colored in places. This rock type forms the most characteristic facies type of the Turonian in Palestine. According to Blake,[47] it is a fine-grained, semi-crystalline rock with the general properties of a lithographic stone. Styolite seams are common. Marble-like, coarse-grained limestone masses, locally pink mottled, form the rock type of the interbedded rudistite zones. They are probably identical with the "? Meleke" in Blake's[48] section from the Jabal Sih. At Khallat et Tayun (southeast of Nazareth) thin, irregularly bedded argillaceous limestones occur intercalated in the Mizi helu. The Mizi helu northeast of Kafr Kanna at some places contains chert nodules bearing foraminifera. Oölites were found in cavities of fossils and in form of small lenses in a fossiliferous Mizi helu bed on the east side of the Jabal Sih.

Fauna.—In the lower complex the following fossils were found:

Miliolidae div. sp.
Alveolina fraasi (Gümbel)
? Trochosmilia sp.
Coelosmilia sp.
Pedinopsis sp.
Holectypus sp.
Echinoid plates and spines
Bryozoa
Praeradiolites maroni Douvillé
Radiolites sp.
Durania sp.
Plicatula cf. *reynesi* Coquand
Nerinea subgigantea Blanckenhorn
N. requieniana d'Orbigny
Iteria sp.
Calcareous algae

Bioherms.—The most abundant fossils are the rudistites, which occur as widespread bioherms and shell breccias. Commonly, there is a single rudistite bioherm, but southeast of Nazareth two separate rudistite zones occur. The lower zone consists of three successive bioherms, of which the middle one was less densely populated. Each bioherm is overlain by re-worked zones, composed principally of rudisite fragments and subordinately of oyster and echinoid remains. In a single instance what appears to be the detrital fan at the outer edge of the lower bioherm zone is exposed. Here (east side of the

[46] Arabic term meaning "the sweet stone."

[47] *The Stratigraphy of Palestine . . . .*, p. 106.

[48] *Ibid.*, p. 39.

Wadi el Haij) the bioherm-bearing zone is replaced by an unusual thick unstratified shell breccia. The upper zone contains only one bioherm, again overlain by a shell breccia. The biotope of the interbioherm areas differs considerably from that of the bioherms proper. With increasing distance from the bioherm edge, rudistites and their shell fragments disappear gradually and are replaced by a faunal assemblage, which consists of the above-listed individual corals, echinoids, pelecypods, and gastropods. The local find of fragmentary bryozoans and calcareous algae in an interbioherm assemblage, in close proximity to the bioherm edge, makes it appear uncertain whether these forms belonged originally to the bioherm or interbioherm fauna. The faunal assemblage of the shell breccias overlying each of the bioherms indicates temporary overlaps of the interbioherm biotopes. These intercalated re-worked zones might be interpreted as indications of differential diastrophic movements.

An alternative explanation may be deduced from two facts: (1) the extreme shallow-water conditions to which the rudistites had been adapted; (2) the limited vertical development of the individual bioherms. It is conceivable that limited vertical growth of the rudistites in such an environment brought the bioherm within the reach of extreme wave action and eventually caused intertidal exposure of the rudistites to the air. A shortening of the food-gathering period due to prolonged closure of the valves may have killed the rudistites, and wave action then destroyed their skeletons. Additional evidence, however, is necessary to decide the issue.

Correlation.—*Alveolina fraasi* was found widely distributed in the area studied in the lower complex, which corresponds stratigraphically to the lower part of the Mizi helu (lower Mizi beds) in the Wadi el Jor at Jerusalem. Specimens collected from the latter locality were originally described by Fraas[49] as "*Nummulites cretacea*,"[50] which may be regarded as an important zone fossil in the region studied, since it occurs in both biotopes. Further study in other regions where the Mizi helu facies extends to the top of the Turonian

[49] *Op. cit.*, pp. 55–56, 83–84.

[50] C. W. Gümbel, "Über zwei jurassische Vorläufer des Foraminiferen-Geschlechs, Nummulina und Orbitolites," *Neues Jahrb. f. Min.* (1872), pp. 241–59. Gümbel, who examined specimens collected by Fraas, called them *Alveolina fraasi*.

sequence will show whether *A. fraasi* is restricted to the lower part of the Turonian or is a facies fossil.

Of the specifically identified forms, *A. fraasi* and *P. maroni* are confined to the Turonian. *Praeradiolites maroni* has been reported only from the Lower Turonian of the Lebanon.[51] The remaining forms are of little value in age determination, owing to their extended stratigraphic range.

B. UPPER COMPLEX

Lithology.—The upper complex is a well-stratified series of pinkish mottled, crystalline limestones. Beds of the Mizi helu type commonly occur intercalated in the basal part of this series. Argillaceous limestones and dolomites are locally developed.

The crystalline limestone beds of the upper complex grade into the chalky limestones of the overlying Santonian deposits.

Fauna.—The upper complex yields the following fossils:

Rhynchonella plicatilis var. *bohemica* Schloenbach
Exogyra columba Lamarck
Pycnodonta vesiculosa Sowerby
Pycnodonta vesicularis Lamarck
Ostrea div. sp.
Fish teeth and bones

A pronounced faunal change marks the upper complex. The Rudistaceae with their faunal associates are replaced by a monotonous fauna of Rhynchonellidae, Ostreidae, and fish remains. Beds with *Ostrea vesicularis* mark commonly the passage zone to the overlying Santonian.

Correlation.—*Rhynchonella plicatilis* var. *bohemica* is confined to the lower part of the upper complex. The species can be regarded as a true zone fossil, as its areal distribution in the Eastern Nazareth Mountains shows. According to N. Schloenbach[52] and H. Scupin,[53] this form occurs in the Lower and Middle Turonian of central Europe and France. The Ostreidae are of little stratigraphic value because of their extended range.

The fossil content, though meager, establishes the upper complex

[51] Douvillé, *loc. cit.*

[52] "Kleine paläontologische Mittheilungen, III. Die Brachiopoden der böhmischen Kreide," *Jahrb. d. k. k. geol. Reichsanst* (Wien), Vol. XVIII (1868).

[53] "Die Löwenberger Kreide und ihre Fauna," *Palaeon. Suppl.*, Vol. VI (1912–13), pp. 246–47.

as Turonian in age. The limited number or forms in both the lower and upper complex which are confined to the Turonian confirms the general observation that the Turonian fauna in Palestine, in contrast to that of Syria and Transjordania, is not well defined.[54]

Consideration of Blake's section of the Jabal Sih shows that both his Upper and Lower Turonian are part of what we have designated "Lower Turonian complex." Picard describes similar beds with *Ostrea* sp. and *Rhynchonella* sp. from the Western Nazareth Mountains.[55] He suggests a Santonian age for these rocks, a correlation based exclusively on lithologic criteria.

Thickness.—The Turonian beds have a total thickness of 20–30 meters.

SENONIAN

GENERAL STATEMENT

The Senonian comprises a sequence of unstratified chalks, marls, chalky and pure limestones with a distinctive fauna. There is no definite boundary toward the underlying upper complex of the Turonian because of the grading facies previously mentioned. The upper boundary is commonly well marked by the appearance of the well-bedded chalky limestone of the "transitional" Cretaceous-Eocene series. On the basis of faunal evidence the Santonian, Campanian, and Maestrichtian stages, as defined in the Senonian key section in the Judean Desert,[56] are recognizable. Difficulties, however, were experienced in attempting to separate the Campanian section from that of the Maestrichtian, owing to the localized occurrence of fossils, facies, and faunal changes and extensive masking of the deposits by "nari."

AREAL DISTRIBUTION

The areal extent of Senonian exposures is shown on the geological map (Fig. 2).

PREVIOUS WORK

A comparison between the accompanying map and the one by Blanckenhorn[57] shows that most of the areas marked by Blancken-

[54] Blanckenhorn, "Geologie Palästinas . . . . ," *op. cit.*, p. 10.

[55] Picard, "Zur Geologie der Kischon-Ebene," *op. cit.*, p. 9.

[56] Picard, *Geological Research in the Judean Desert* (Jerusalem, 1931), p. 1.

[57] "Kurzer Abriss . . . . ," *op. cit.*, map.

horn as Senonian are Eocene in age. Blanckenhorn's view was adopted in the later maps of Fohs,[58] Range,[59] and Dubertret.[60] In a composite section of the sequence at the Jabal Sih, Blake[61] recognizes the Santonian, Campanian, and Maestrichtian stages of the Senonian. Among "good examples" of the Maestrichtian, Blake[62] cites an outcrop north of Nazareth which, according to the following discussion, exposes deposits of the transitional Cretaceous-Eocene to Lower Eocene series. The age of another occurrence of the same series at the east side of Jabal Sih[63] is given as "Lower Middle Eocene or Maestrichtian."

A. SANTONIAN (CONIACIAN)

Lithology.—The Lower Senonian consists of massive white chalky limestones, which in places are pink mottled. This rock type, the so-called "Kakuhle," represents the characteristic Santonian facies of Palestine. On the basis of its uniform lithology throughout the area studied, the Santonian can be separated even where it does not carry megascopic fossils.

Thickness.—The thickness of the Santonian is 6–10 meters.

Fauna.—The following fossils were found only in the northern belt of outcrops, near Kafr Kanna and Lubya:

Protocardia sp.
Ostrea sp.
Dentalium cretaceum Conrad
Mortoniceras cf. *quattuornodosum* var. *planulata* Lasswitz
Gauthiericeras austinense F. Roemer
Gauthiericeras austinense var. *minima* Lasswitz

Correlation.—According to Blanckenhorn,[64] *M. quattuornodosum* and *G. austinense* are characteristic forms of the Santonian of Palestine. They had previously been described from the Austin chalk of Texas. The impression of a large ammonite mold, observed at the

[58] *Op. cit.*, map.

[59] "Wissenschaftliche Ergebnisse einer geologischen Forschungsreise nach Palästina im Frühjahr 1928," *op. cit.*, sketch map.

[60] *Op. cit.*, map.

[61] *The Stratigraphy of Palestine* . . . . , p. 39.

[62] *Ibid.*, p. 46.

[63] *Ibid.*, p. 39.

[64] "Geologie Palästinas . . . . ," *op. cit.*, p. 11.

fossil occurrence of Kafr Kanna, indicated the presence of *Texanites quinquenodosus*, the common zone fossil of the Santonian of Palestine. Thus the Santonian may be said to be both lithologically and faunally well marked in the Eastern Nazareth Mountains.

B. CAMPANIAN

Lithology.—In contrast to the uniform lithology of the Santonian, the lower part of the Campanian sediments varies considerably in lithology. Iron-stained, foraminiferal marls, hard chalky limestones, and soft white chalks form the basal Campanian in the few exposures observed. A lateral tracing of the different facies was impossible, since most of the rocks are hidden by a nari cover up to 2 meters in thickness. The overlying deposits consist uniformly of soft chalks which locally are highly bituminous. A dark-brown flint bed is intercalated in these chalks throughout the area studied. Locally, in the Wadi Ramiq, however, three flint beds were found, underlain by an oyster bed, which consists of thin shells of *P. vesicularis*. Oyster beds and the parallel orientation of shells of *D. cretaceum* indicate shallow-water conditions, with the sea floor temporarily, at least, above wave base.

Fauna.—The lower Campanian deposits are highly fossiliferous. The following fossils were collected predominantly from this zone at Kafr Kanna, in the Tur'an plain (southern border) in the Wadi Umra Jubeil, at Arab S'Beih, and in the vicinity of Nazareth:

Magas cf. *geinitzi* Schloenbach
M. sp.
Leda perdita Conrad
Arca fabiformis Conrad
A. (*Grammatodon*) *parallelus* Conrad
Astarte undulosa Conrad
Crassatella cf. *falconieri* Lartet
C. rothi Fraas
C. cf. *larteti* Picard
Liopista sp.
Meretrix andersoni B. Newton
Lucina cf. *dachelensis* Wanner
Corbula cf. *paracrassa* Wade
Inoceramus sp.
Pecten delumbris Conrad
Pycnodonta vesicularis Lamarck
Dentalium cretaceum Conrad
D. octocastatum Fraas
Natica sp.
Scalaria sp.
Turritella maussi Lartet
Fusus sp.
Volutilithes sp.
Ammonites sp. ind.
Ancycloceras sp.
Scaphites sp.
Baculites sp.
Fish remains

The faunal character is Mediterranean with local eastern Mediterranean features.

Correlation.—With the exception of the brachiopods cited, which represent a new faunal element, most of the forms are commonly known from the Upper Senonian of Palestine and Egypt.

C. MAESTRICHTIAN

Lithology.—In the upper part of the Senonian sequence changing facies conditions recur. Commonly the Maestrichtian is represented by a white (in places, hard) chalky limestone, which, because of the abundance of *Baculites* casts, is characterized as "*Baculites* facies." At some places the bituminous chalk facies of the Upper Campanian extends without visible change up to the top of the Maestrichtian (Wadi Ramiq). In this facies the bitumen content increases locally compared with that in the Campanian (Bir Kapshane). Iron-stained foraminiferal marls, lithologically identical with the Lower Campanian facies, form the uppermost Maestrichtian near Sajara and north of Er Reina. Locally, flint beds were observed in the Maestrichtian.

Bone-bed-like concentrations of fish remains, associated with granules of phosphate, occur as small lenses in the chalky limestones. Fragments of bones, scales, and teeth of fishes form the bulk of the constituents. Coprolites and fragments of oyster shells are less common. The interstitial groundmass is predominantly composed of lenticules and oölites of phosphate and, to a minor extent, of foraminifera of the size of the phosphate granules.

Facies discussion.—The sporadic occurrence of both flint and phosphate likewise characterizes the Maestrichtian in the standard area, the Judean Desert. There flint occurs in numerous compact beds and bands of nodules, and the phosphate is found in one or two thick zones.[65] A comparison of the Maestrichtian sections of the Judean Desert, of the Wadi Farah region[66] (Samaria), and of the area studied demonstrates a progressive decrease in the thickness of the Maestrichtian deposits and, more especially, of the flint and

[65] Picard, *Geological Researches . . . .*, p. 31.

[66] Picard, "Zur Geologie des Gebietes zwischen Gilboa und Wadi Fara," *Centralbl. f. Min. Geol. Pal.*, Abt. B (1934), p. 25.

phosphate components from the Judean Desert to the Eastern Nazareth Mountains. The same observation was made by M. Avnimelech[67] by comparing the Maestrichtian of the Judean Desert with that of the western flank of the Judean arch.

There is the question as to what the different facies indicate. Picard pointed out that the formation of the phosphate and flint deposits, as well as the presence of abundant semisessile faunal elements, indicates extreme shallow-water conditions of the Maestrichtian facies in the Judean desert area. There are ample data to show that the Maestrichtian sea of the eastern Nazareth area was also shallow. Such evidence may be seen in the sporadic occurrence of phosphatic bone-bed lenses, foraminiferous sands, the abundance of Ostreidae, the local concentrations of uniform-sized specimens of genus *Baculites*, and the parallel orientation of elongated shells of gastropods, scaphopods, and fish bones. In contrast to the extreme shallow-water conditions indicated for the entire Judean desert area, only occasional, local indications that the sea floor was above wave base are observable in the area studied. The water depth, there, may have been near the wave base.

The amount of uplift responsible for the shallowing of the sea in Maestrichtian time, compared with the Campanian, was therefore much greater in the Judean desert area than in the region studied. These intensified diastrophic movements at the end of Cretaceous time resulted in the complete emergence of parts of Judea and Transjordania[68] as an island, in which the Judean desert area was centrally situated. In the area studied, however, sedimentation went on, with probably minor local exceptions, up to the end of the Middle Eocene. The facies conditions in the area were, therefore, more typical of the epicontinental Maestrichtian sea than those of the standard area, the Judean Desert, which were rather local in character.

Fauna.—The Maestrichtian sediments are, like those of the Campanian, rich in fossils. The *Baculites* facies may be regarded as diag-

[67] "Etudes géologiques dans la région de la Shéphélah en Palestine" (thèse, Grenoble: Allier Père et Fils, 1936), pp. 43–45.

[68] The extent of the island is shown on the sketch map by Blake in "Old Shore Lines of Palestine," *op. cit.*, p. 72.

nostic for the Maestrichtian, just as is the case in the Judean desert area.

The following fossils were collected in the vicinity of Nazareth, Er Reina, Kafr Kanna, and Esh Shajara:

Cyclocyathus(?) sp. Felix
Serpula aff. *subtorquata* Münster*
Leda perdita Conrad
Grammatodon(?) *Arca parallelus* Conrad
Arca fabiformis Conrad
Astarte undulosa Conrad
Crassatella cf. *rothi* Fraas
C. falconieri Lartet
Liopista sp.
Lucina safedensis Conrad
L. dachelensis Wanner†
L. saharica Fraas*
Meretrix andersoni B. Newton
Cytherea aff. *rohlfsi* Fraas*
Corbula cf. *paracrassa* Wade
Inoceramus regularis d'Orbigny
Pecten nilssoni Goldfuss
Pycnodonta vesicularis Lamarck
Pecten spatulatus (Ravn)
Ostrea sp. (cf. *O. nicaisei* Coquand)
Dentalium octocostatum Fraas
D. cretaceum Conrad
D. sp.
Natica aff. *farafrensis* Wanner†
Turritella maussi Lartet
T. reyi var. *seetzeni* emend. Blanckenhorn†
Mesalia foucheri Pervinquiere*
Cerithium aff. *pseudoclathratum* d'Orbigny*
Helicaulix (*Aporrhais*) cf. *stenoptera* Goldfuss*
Aporrhais sp. nov.(?)
Ammonites sp.
Hamites sp.
Heteroceras sp.
Baculites sp.†
Pseudochloenbachia sp.
Ancycloceras(?) *safedensis* Conrad
Scaphites sp.
Aptychus sp.
Corax pristodontus Agassiz*
Fish remains

* Known from the Maestrichtian of other countries.
† Known from the Maestrichtian of the Judean Desert according to Picard.

The fauna is readily distinguishable from that of the Campanian, especially where the *Baculites facies* is developed.

Thickness.—The Maestrichtian and Campanian sediments have a total thickness of 60–80 meters. No figures for the thickness of each stage were obtainable, as pointed out above. From partial sections it can be estimated, however, that the thickness of the Maestrichtian is less than 10 meters.

Danian(?).—The uppermost 1–2 meters of the Maestrichtian sediments are free of megascopic fossils. Since the lithology does not

change in this unfossiliferous part of the sequence, these rocks were provisionally included in the Maestrichtian. They may represent, however, the equivalent of the Danian.

CRETACEOUS HISTORY

The geologic history of the Eastern Nazareth Mountain area can be traced back to the Vraconian stage. At this time the Tethys Sea invaded the Nazareth area, which, geographically considered, is a marginal part of the Arabian shield. Although no earlier deposits are exposed, it can be concluded from the approximate course of the shorelines of the Lower Cretaceous sea[69] that there were two earlier marine invasions in Aptian and early Albian time followed shortly by retreat. In contrast to these oscillations, the submergence in Vraconian time was the initial phase of a continuous widespread transgression, the so-called "Upper Cretaceous" or "Cenomanian transgression," recognized in many parts of the world.

The Vraconian sea advanced from the west and was, at the beginning, very muddy, owing to the proximity of the shoreline to the east (about 40 km.). Toward the end of the stage a clearing of the sea took place. Thus, sedimentation in early Cenomanian time was practically limited to chemical and organic precipitates, calcareous sediments which have been altered predominantly to dolomitic rocks, possibly diagenetically. At this time the progressive subsidence of this part of the shelf since Vraconian time reached its maximum, and the sea was relatively deep and characterized by local sponge reefs. In late Cenomanian time recurrent movements caused a shallowing of the sea, and there are local indications that the sea floor was above wave base. These shallow-water conditions prevailed through Turonian time. They favored the development of extensive rudistite reefs in the clear water of the first part of the Turonian.

The limestones of the Turonian grade upward into chalky limestones of the Santonian stage, at the beginning of Senonian time, and into chalks and marls of the following Campanian and Maestrichtian stages. These fine calcareous foraminiferal muds appear

[69] *Ibid.*, p. 70, Fig. 1; also p. 72.

to represent far-offshore deposition on the shelf. During Santonian time the sea seems to have deepened slightly, but the local concentration of Ostreidae shows that it was still relatively shallow. The occurrence in the Tethys Sea of a few ammonite species of the genera *Mortoniceras* and *Gauthiericeras*, characteristic of the far-distant Texas areas, appears remarkable. The following stages of the Senonian, the Campanian, and the Maestrichtian are characterized by an abundance of organic remains (especially of mollusks), which is typical for the entire Palestinian part of the shelf. A progressive shallowing of the sea during both stages is clearly evident, with the climax in late Maestrichtian time. This progressive shallowing, as evidenced in the region studied, was only a minor reflection of those uplift movements whose center was farther southeast between Nablus and Beersheba on the Palestinian side and in the northeastern adjacent area of Transjordan. The continuance of these movements finally resulted in complete emergence of this area as an island in the early Eocene sea. In the region studied, however, continuous sedimentation seems to have prevailed from late Maestrichtian to early Eocene time, though local diastems may be present.

THE ORIGIN OF SPITS, BARS, AND RELATED STRUCTURES[1]

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ABSTRACT

Building and lengthening of the superaqueous portion of a spit or bar is caused by the transportation of material from the shore toward the distal end of the structure by the swash and backwash when the waves break obliquely on the shore. If the water surface has a constant elevation, the submerged part of a spit or bar is not brought above the surface by waves moving in a direction perpendicular to its axis. Such wave action results only in widening the submerged ridge.

Hooks are caused by the work of refracted waves which cause transportation of material across the end of the spit. A change of wind direction or the presence of tidal or hydraulic currents is not necessary to their building, although if such currents are present they may aid the process. Looped bars are built by refracted waves in the same way as are hooks. However, in the looped bar the process operates more uniformly and continues until the bar rejoins the shore. The windward end of a looped bar originates in shallow water on a gently sloping bottom. The location of the point at which the return to shore begins depends on the depth of water, the amount of sediment being carried by beach drifting, and the energy of the waves.

Extensive observations indicate that the lengthening of spits and bars takes place only when the waves move in a direction such that they cause the swash and backwash to move material along the shore toward the end of the spit. Waves from any other direction cause it to be widened and shortened.

INTRODUCTION

A "spit" is a ridge or embankment of sediment attached to the land at one end and terminating in open water at the other. It is younger than the land mass to which it is attached. The crest of the spit from the land outward for some distance rises above the water.

A "bar" is a completed or extended spit which encloses, or nearly encloses, a portion of the water body into which it extends. It may be attached only at one end, or it may be the result of two spits building from opposite directions. If such a bar extends across a bay it is called a "bay bar." If it departs from a relatively straight shoreline and then swings back it is a "looped bar."

An "offshore bar" is a ridge or embankment of sediment lying approximately parallel to a shoreline and reaching above the water. It is not attached to the shore at either end. Like the spit and the bar, it is younger than the land body along which it lies.

[1] With the support of a grant from the American Philosophical Society.

Any of these structures may vary greatly in size and age. Some spits are only a few feet long and have a life of only a few hours. Examples of these can be seen where small streams empty along sandy shores which are acted upon by waves produced under the influence of variable winds. Others may be large and have a long life, such as Sandy Hook or Cape Cod. Likewise, bars may shut off lagoons a few feet across or may close the entrance to great bays large enough to form important harbors. Unlike spits and bars, offshore bars of small size are not very common. The offshore bars usually described in print are such as those along the coast of New Jersey or the great offshore bars of the Gulf Coast or the Carolinas.

In this article the processes by which spits and bars are formed will first be considered. Offshore bars will be discussed only with regard to their possible relation to spits and bars.

Our geological textbooks give a somewhat indefinite and hazy explanation of the methods of formation of spits and bars. The most common statement is that they are the result of transportation by the longshore currents, but usually no attempt is made to give a complete picture of the process. Of sixteen modern geological textbooks of college grade which were examined, five stated that spits and bars are built by shore currents, ten that they are the result of shore currents combined with the work of the waves as they encounter the submerged part of the ridge, and one that they are built by the shore current which is sometimes helped by "wave wash from opposite directions."

D. W. Johnson[2] gives an excellent summation of the literature of spits and bars which is still fairly complete. The earlier studies of spits, bars, and related structures were confined chiefly to descriptions of their physical forms. Attempts at explanation of origin were largely deductive. A good example of this is F. P. Gulliver's work.[3] Little was known at that time of the processes by which sediments are transported in water bodies, and very few systematic subaqueous observations had been attempted. Gulliver recognizes this weakness in his own work, for he states that "its confirmation, extension, or re-

[2] *Shore Processes and Shoreline Development* (New York: John Wiley & Sons, Inc., 1919), pp. 333–39.

[3] "Cuspate Forelands," *Bull. Geol. Soc. Amer.*, Vol. VII (1896), pp. 399–422.

jection awaits the local observer in the field." In field work emphasis was at first placed largely on the study of fossil structures. An outstanding example of this is G. K. Gilbert's classical work.[4]

Later work by various investigators made it clear that the energy of longshore currents alone is hardly sufficient in itself to explain the formation of all spits and bars, especially those containing coarse material. Also, it was soon realized that spits and bars are numerous along the shores of small bodies of water where eddy and tidal currents are either absent or negligible. A few years after the publication of Gilbert's studies, Tarr, Woodman, and Wilson, each as the result of independent work, suggested wave action as a possible cause of the formation of spits and related structures. However, they failed to make any adequate investigations of processes of sediment transportation. Consequently, their explanations do not give a clear picture of the exact processes by which the structures are built.

Johnson, in his excellent book,[5] appears clearly to recognize the incompleteness of the explanations that have been offered and emphasizes the importance of beach drifting and wave-formed currents in the formation of shore structures. He gives an analysis of the process of beach drifting[6] and seems to be the first to have seen the great importance of the swash and backwash in the transportation of sediment. However, as will appear later, even he probably failed to appreciate fully the extremely important role played by the beach-drifting process.

Since spits and bars extend above the water, any genetic account of their origin must explain the building of the exposed portion as well as that part below the surface. This is the weakest point of the explanation usually given in the textbooks. It is easy to understand how currents, if strong enough to pick up sediment, may again deposit it under water in the form of ridges; but it is evidently impossible for such currents operating entirely within the water to deposit above its surface.

The necessity for explaining this point seems to be thoroughly understood by Johnson, who gives an extended discussion of the subject in describing the building of offshore bars.[7] However, he follows

[4] "Lake Bonneville," *U.S. Geol. Surv. Mono. I* (1890).

[5] *Op. cit.*, pp. 335–39. [6] *Ibid.*, p. 94. [7] *Ibid.*, pp. 356–70.

De Beaumont, Shaler, and Davis in assuming that the action of the waves is capable of building a submerged offshore bar above the surface of a body of water having a constant elevation. The writer has not had access to De Beaumont's original article, but Shaler, Davis, and Johnson[8] give rather complete and essentially identical descriptions of the building of offshore bars and of the process by which they are supposed to be brought above the water, although none of them except Johnson appears to have seen the process in operation. He says:

> The writer has seen a very perfect miniature off-shore bar formed in a few hours by waves raised on the surface of a small lake at Lakehurst, New Jersey, during a fresh breeze. The bar was a few inches in width, and separated a shallow lagoon one or two feet broad from the gently sloping shore which it paralleled for some yards.[9]

However, there is a possibility that even as experienced an observer as Johnson may have been deceived in this instance. The level of the water on the lee side of a small lake is often raised several inches during strong onshore winds. At such times it is not unusual for small subaqueous ridges of sediment to be built up by longshore currents nearly to the surface of the water and then to be driven shoreward by the waves of translation across them. Such a ridge may appear above the water surface as the wind goes down, as a result of the fall in the water-level. Such a process would be entirely different from that of building the ridge above the water surface by the work of the waves.

But extensive inquiry among commercial fishermen, members of the coast guard, and others who have lived along the shore has failed to discover anyone who has ever seen an offshore deposit of sand or mud built above the surface except when the water surface has been lowered or the growth of weeds has checked the water movements sufficiently to cause deposition. It seems probable that, if offshore bars can be formed by the work of the waves, they should be numerous along the shores of some parts of the Great Lakes because of the many subaqueous sand ridges or "balls" found in many places in the Great Lakes and especially along the east side of Lake Michigan.

[8] *Ibid.*, pp. 365–66.

[9] *Ibid.*, p. 388.

According to my studies,[10] it is impossible for a subaqueous ridge which is not connected with the shore to be built above a water surface of constant level by wave action. As soon as such a ridge is brought near the surface the waves of oscillation break against it. This causes waves of translation which carry the sediment across its top, and its further upward growth is prevented. Thus the material on the top of the ridge is planed off and deposited on the lee side, and the ridge migrates as a subaqueous dune.[11]

Johnson appears to recognize clearly the necessity of explaining how spits and bars are brought above the water surface, but he seems to accept completely the hypothesis of wave-built ridges as it has been applied to offshore bars and attempts to apply it also to spits and bars. He says: "The seaward side of the narrow embankment is acted upon by the ocean waves, which build its crest above normal sea-level and establish a profile of equilibrium, similar to that of an ordinary beach."[12] He also says:

> The super-aqueous portion [of the spit] owes its height primarily to the waves, but in the case of sand spits wind action may locally raise the level a number of feet by forming dunes. Disregarding the disturbing effect of the wind, the height of a spit will depend upon the exposure to wave action; big waves will cast the debris many feet above mean water level, while small waves will raise the surface but slightly above the lake or sea.[13]

Again, in discussing compound spits, he says:

> Large quantities of debris, borne by a current which departs from the former shoreline and advances into open water, must be built into an embankment which elongates rapidly in the direction of current advance. Waves raise the surface of the new embankment into a beach ridge; and by repetitions of this process there are formed successive beach ridges separated by lagoons of considerable breadth.[14]

In describing the building of bars Johnson says:

> There is, however, an entirely different process by which bars, indistinguishable in surface form from those developed from growing spits, may be pro-

[10] Evans, "The Low and Ball of the Eastern Shore of Lake Michigan," *Jour. Geol.*, Vol. XLVIII (1940), p. 497.

[11] *Ibid.*, pp. 498–99. The ridges of the low and ball formed by the plunging breakers are never brought into this upper zone of migration except by accident, but there are numerous ridges nearer shore, which are formed in other ways, that do commonly migrate shoreward in this way.

[12] *Op. cit.*, p. 287. [13] *Ibid.*, pp. 295–97. [14] *Ibid.*, p. 298.

duced. Waves entering shallow water may break before reaching the coast, and cast up the bottom debris into a narrow ridge, in the manner discussed more fully in connection with "Offshore Bars." The irregular bottom of a typical young shoreline of submergence is usually highly unfavorable to this process; but whenever the initial form or later deposition does give a fairly uniform slope to the bottom near the shore, wave action may produce a bar independently of longshore transportation. Such a bar may form a short distance offshore and be driven in until the portion opposite a headland becomes a headland beach, and the portion opposite the bay remains a typical bay bar extending from headland to headland and nearly or quite closing the bay mouth; or the waves may construct the bar just at the mouth of the bay in the first place; or they may break on the gently sloping bottom well within the bay and produce a bar near the middle or even near the head of the bay. It is possible that some supposed sandspits are really the beginnings of, or last remnants of, bars formed in this manner.[15]

He frequently refers, however, to the importance of "beach drifting" and "wave currents." For example, he says:

There can be no doubt that wave currents and the associated longshore beach drifting play a very important role in the formation of various types of beaches, spits, bars, tombolos, and forelands.[16]

But nowhere does he speak of beach drifting as a direct factor in building the structures above the water surface, while he does, as indicated above, repeatedly refer to the efficiency of breaking waves in bringing subaqueous embankments above the water surface. Therefore, it seems evident that Johnson's hypothesis of spit- and bar-building includes two processes: first, the building of subaqueous embankments by longshore currents and, second, the elevation of these embankments above the water by direct wave action in a way similar to that in which beach ridges are thrown up.

Since direct wave work is incapable of bringing offshore subaqueous embankments above a water surface of constant elevation, it is not probable that spits and bars are built above the water by the processes heretofore supposed. This study was undertaken, therefore, for the purpose of determining the exact processes involved in the building of such structures.

THE REGIONS STUDIED

The first observations were directed toward discovering, if possible, any instance where breaking waves were effective in building the

[15] *Ibid.*, p. 301. [16] *Ibid.*, p. 334.

submerged end of the spit above water. It is evident that if subaqueous embankments can be brought above the water by the direct action of the waves, the most effective direction of wave travel is perpendicular to the axis of the submerged end of the spit, as at *A* in Figure 1. With the waves as indicated at *B* the conditions would be less favorable than at *A*, but there should still be some building, since one component of the wave direction at *B* is perpendicular to the axis of the spit. With the wind as shown at *C* we would not expect an extension of the spit, but there should be no decrease in its length. Should the wind be from the direction shown at *D*, the observations would be indeterminate, since in this case it would be impossible to distinguish between the effects of wave-building and beach drifting. If wave-building is the effective agent by which embankments are brought above the water, there might well be built occasionally, with the wind as shown at *A* or *B*, small islands of sediment along the crest of the ridge because of inequalities in wave energy and in the amount of sediment available. These islands, if later joined, would become a part of the spit built above the water.

For this study three spits were selected. Two of these are permanent, but one is temporary, in that it alternately grows and is destroyed, depending on the wind direction. One of the two permanent spits is located on the south side of Blue Lake a few miles northeast of Whitehall, Michigan. It points east and is at the end of a small cape which partly encloses a bay about 200 feet across. The spit was first measured on June 12, 1941, following several days of fairly strong northeast and east winds. At that time it was $27\frac{1}{2}$ feet long and 18 feet wide. Its outer end was almost semicircular in shape and had the appearance of having been blunted and cliffed by the recent waves. On June 17, following three days of strong northwest winds, the spit had been extended east by south 4 feet and was $31\frac{1}{2}$ feet long (see Fig. 2). A strong beach ridge had developed along the north side of the structure, evidencing that the material added to its outer end had come by the process of beach drifting from the sandy shore to the west. Although the wind had dropped somewhat at the time of the visit, transportation and building were still going on, the waves on reaching the end of the spit being refracted in the shallow water and moving around its end toward the bay. As a consequence, beach

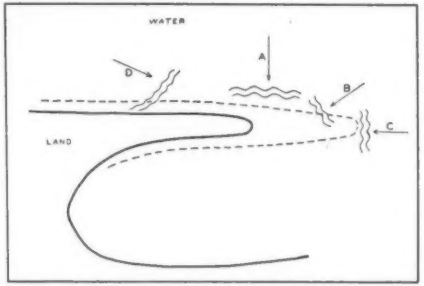

FIG. 1.—Action of waves in spit-building

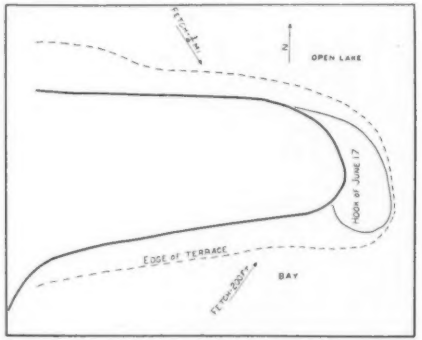

FIG. 2.—Blue Lake spit

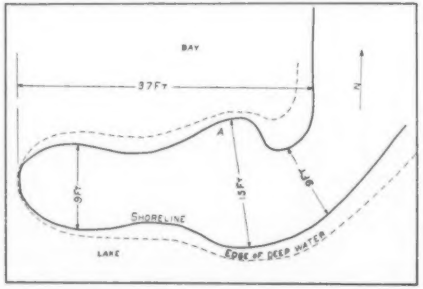

FIG. 3.—Duck Lake spit, June 17, 1941

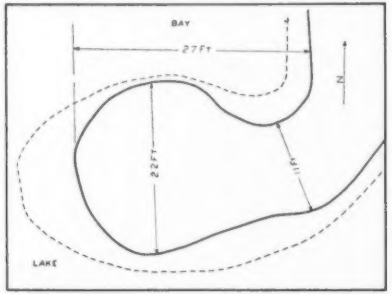

FIG. 4.—Duck Lake spit, August 6, 1941

drifting across the end of the spit was forming a small hook pointing south. For some days following, the waves were gentle, and little change took place. During the night of July 1 the winds were fresh west-northwest and by mid-forenoon of July 2 a hook at the end of the spit was forming. Some cliffing had occurred on the north side and across the end of the point, and the spit was now shortened to 29 feet. On July 19 strong northwest winds were again blowing, and the spit was again lengthened to nearly 31 feet. The wind was a little more northerly than on June 17, and the waves were being refracted more strongly around the end of the spit and were tending to broaden it. Thus this spit was lengthened by material brought along the beach ridge during northwest winds and was shortened and rounded with a strong tendency toward hook formation when the wind was north or northeast.

The second permanent spit is a deposit of sand on the north side of Duck Lake about 2 miles south of White Lake Harbor entrance. It is an extension of a geologically recent sand point which partly separates the west arm of Duck Lake from the main body of water. The first visit to this structure on June 17 followed a considerable period of strong northeast and east winds, and its general appearance and relation to the shore was as shown in Figure 3. There was a strong beach ridge along its south side, and it had every appearance of having made a recent rapid growth to the west. The end terminated abruptly in relatively deep water. The material of the spit clearly had come from the abundant sand to the northeast. Several ensuing days of gentle to moderate southwest winds with a fetch of about one-fourth of a mile shortened the spit from 37 to $33\frac{1}{2}$ feet, increased its width at the end from 9 to 15 feet, and turned it more toward the north shore. There was now a subaqueous shelf 3 feet wide at the end and along the south side of the spit where none was present on June 17. At *A*, erosion was taking place and material was moving eastward toward shore. By July 10, after a further period of fairly strong westerly winds, the shelf at the end of the spit had widened to 5 feet, and the length of the spit above water measured only 31 feet. The shelf on the south side had widened to 8 or 9 feet. On the north side at *A* the shoreline was now straight. Here, again, is evidence that winds which cause beach drifting along the axis of the spit from

the land to which it is attached cause it to grow, while winds from the other directions destroy it. By August 6, after a long period of southwest, west, and northwest winds, it had the form and dimensions shown in Figure 4. At this time the winds were moderate from the west, and material was moving eastward by beach drifting on both sides of the spit.

The third study was made on the west side of Silver Lake. Here the Little Point Sable dunes are moving eastward and slowly filling the lake. This sand is shaped by the waves into a shallow, gently

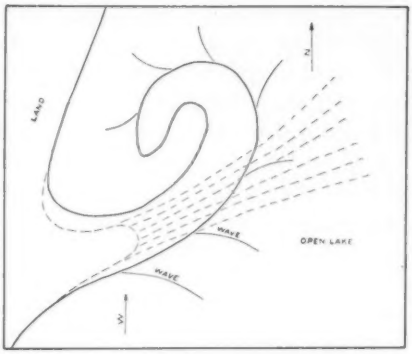

FIG. 5.—Hooked spit, Silver Lake

sloping underwater terrace, and the shoreline is being re-worked with every change of wind. The terrace is so shallow and has such a gentle slope that numerous cusps form, some of which extend almost straight out from shore and terminate in long subaqueous ridges that often become subaqueous migrating dunes.[17] The shoreline here lies north by east. On June 12, with the wind east-southeast, one of these low ridges attached to the shore at its south end was being driven toward shore. The waves were 3–4 inches high, and the water was very shallow on the ridge. By the next day the wind had shifted to the south, the subaqueous ridge had become more nearly parallel with the shore, and a spit had formed at the shoreward end of the

[17] Evans, "Mass Transportation of Sediments on Subaqueous Terraces," *Jour. Geol.*, Vol. XLVII (1939), pp. 327–28.

ridge. The part projecting above the water was 8 feet long and 28 inches wide (see Fig. 5). Beyond where it became submerged, it gradually diverged from shore as the water became deeper, and at about 60 feet it could no longer be distinguished as a ridge. By June 21, with continued south and southeast winds, the spit had become 25 feet long and 4 feet wide, and a hook had formed that turned directly back into the little bay on the landward side of the spit (Fig. 5). This hook was not the result of currents caused by change in wind direction but was the result of beach drifting caused by refraction and diffraction of waves coming from the south and southeast. This process of hook formation will be discussed in greater detail under the subject of "looped bars."

Streams which enter Lake Michigan across the broad sandy beaches of the east side of the lake have their outlets shifted to the south or the north as the direction of the wind changes. When such shifting occurs, it is always accompanied by the building of spits which have their shore attachment on the side from which the waves come. The growth is always in the direction of beach drifting, and the more material carried forward by the process the more rapidly the spit grows. When the winds happen to be almost directly onshore, the process is very slow, and spits already formed are sometimes shortened. However, this depends on the relation between the sand-transporting energy of the waves and that of the stream. In no case is the mouth of the stream ever closed simultaneously throughout its width by a ridge thrown up by the waves. The outlets at Duck Lake and at Flower Creek were especially kept under observation during the summer of 1941; and these, with others, have been under occasional observation for a much longer time. At no time was there observed a tendency toward spit-building by any process except that of shore drifting, although the rapidity of the process was greatly increased by the amount of sand brought down by the stream.

Thus it seems clearly indicated that spits and bars increase in length above the water only when the waves move in such a direction that material is moved in the swash and backwash outward from the attached end. Waves coming from the other direction or perpendicular to the axis of the structure tend to shorten it. However, beach

drifting along the side of a spit does not necessarily lengthen the spit. Because of wave refraction in the shallow water the sand may be carried around the end of the spit and thus tend to form a hook (Fig. 5). The spit is increased in length only when the amount of sediment carried along its windward side by beach drifting is more than can be forced around its end by the work of the refracted waves. The processes of transportation involved are largely those which have been described by Johnson[18] under the head of "wave currents." Of course, wind-caused longshore currents are also usually present. These aid in the process by building up the bottom and thus shallowing it over the submerged end of the spit.

If Johnson did not accept so completely the theories of De Beaumont in explaining the building of subaqueous embankments above the water, he would seem to be in complete accord with the above statements. He says:

> I can say that I have found many forelands and embankments which seemed to me demonstrably due, principally if not wholly, to wave currents; but none which seemed undoubtedly the product of other types of currents. I am therefore inclined to believe that wave currents have played the most important part in the construction of all the sandy forelands and embankments of our coasts.[19]

Elsewhere he accepts beach drifting as an important process in the formation of beaches, spits, and bars.

An interesting corollary of the above-described process of spit formation is that the building of hooks is a normal accompaniment of spit-building, and in accounting for them it is not at all necessary to postulate tidal or hydraulic currents across the end of the spit. The shore drifting which accompanies wave refraction is entirely competent to form hooks, and it is not necessary to postulate any other change of wave or current direction to explain the process. The factors involved in hook formation under these conditions are wind direction and velocity, slope of bottom, and amount, size, and texture of material available.

The observations described above were made on relatively small structures for convenience and economy of time in doing the work. With large structures it requires a much longer period of time for the waves and currents to produce easily measured results. However,

[18] *Op. cit.*, pp. 90–106. [19] *Ibid.*, p. 337.

effects in the two cases are the same. It is well known that, as waves approach a shore at an acute angle, refraction may cause them to change direction by several degrees, and the more gradual the slope of the bottom the greater is the change. Also, waves coming into a bay between headlands have their courses changed by refraction, so that as they approach the beach they are nearly parallel with the shoreline. This is discussed by Johnson[20] in explaining the concentration of waves in the erosion of headlands. The same effect can be observed where the entrance to a harbor is so situated that waves drive directly into it during a storm. In discussing this action Jeans says: "In the same way sea-waves which fall on the entrance to a harbour do not travel in a straight line across the harbour but bend round the edges of the breakwater, and make the whole surface of the water in the harbour rough."[21]

In studying the building of hooks or recurved spits along ocean shores it is difficult to distinguish between the work of the refracted waves and that of tidal currents; but in the Great Lakes tides are not a factor. Presque Isle is a compound recurved spit which forms the harbor at Erie, Pennsylvania. At its eastern end the process of hook-building is still active. The records of the United States Weather Bureau show that the prevailing winds at Erie are from the southwest, west, and south; and the winds of maximum velocity are also from those directions. There are occasional winds from the north or east, but their effect is not great. The growth of Presque Isle has been from southwest to northeast, and its eastward progression has been marked by the building of a series of hooks across its eastern end. The last of these extensions had its beginning about 1866 and is still building. It is 3,000 feet long and about 700 feet wide. The last 1,000 feet of the south end of the hook is itself a series of hooks whose axes lie approximately northeast-southwest. The testimony of several employees of the park indicates that both the principal hook and the smaller ones at its south end have been built by waves that refracted around the end of the structure in the manner shown in Figure 5.

[20] *Ibid.*, pp. 74–76.

[21] Sir James Jeans, *The Mysterious Universe* (London: Macmillan & Co., 1934), pp. 42–43.

Obviously, the same principles apply to the formation of cuspate bars and looped bars, since a spit which has its distal end turned back until it again joins the shore constitutes a looped bar. Thus it is not at all necessary to postulate two currents circling in opposite directions in the formation of cuspate bars or cuspate forelands. Johnson[22] agrees with this and also says[23] that cuspate bars sometimes form as the result of a spit turning back toward the shore. He does not discuss the process by which it is brought about, and in his discussion of the formation of hooks[24] he seems to place chief emphasis on currents running athwart the end of the spit. Of course, it is entirely possible that cuspate forelands may sometimes form, as has been stated frequently, by two converging spits growing out from the shore; but those which I have observed in the process of building have been recurved spits caused entirely by wave refraction. They are most numerous in lakes having broad, shallow sandy terraces. Along the shore of Silver Lake near Little Point Sable, Michigan, several of them can usually be seen either completed or in the process of formation. They seem to develop best with a wind blowing at an angle of 30°–45° with the shoreline. Studies of looped bars, both completed and in the process of formation on Silver Lake, as well as a consideration of certain fossil looped bars on the shores of Crystal Lake near Frankfort, Michigan, indicate that this is the usual method of formation.

For several days during the latter part of July, 1941, winds from the south over Silver Lake were fairly strong. On the west side of the lake, where the beach is of pure dune sand, a series of five looped bars formed. The shoreline here lies about north-northeast. The bars started as spits building toward the northeast and were later swung back to the shore by the refracted waves. Three of them were complete by July 29, but two had not completely joined the shore at the north end.

Figure 6 is a diagram of one of these looped bars and is typical of them all. A spit began at *A*, and the distal end of it passed beneath the water as a subaqueous ridge as shown at *B*. It continued to build out above the water until it reached a depth such that material brought forward by beach drifting was carried around the end by the

[22] *Op. cit.*, p. 337. [23] *Ibid.*, p. 318. [24] *Ibid.*, p. 289.

refracted waves and a hook began to form at *C* and to grow toward the shore. The refraction of the waves was increased by the continued broadening of the subaqueous dune at *B*. At *D*, to the lee of this subaqueous ridge, the water was deeper. Here the waves, which were partly translation waves across *B*, reformed into oscillation waves and continued the beach drifting along *C* until it joined the shore. In all five bars the ridges became somewhat narrower and lower from south to north.

A good example of combined spit and hooked-bar formation was seen at Old Mission Point on the east arm of Grand Traverse Bay. On July 12, 1941, its extreme tip to the south-southeast reached out about 200 feet. Below the 40-foot bluff here the gravel and sand beach extends around the point (see Fig. 7). The bay lies to the south and southwest of the point. The longest fetch and probably the strongest winds to reach the point are from the north and northeast. Sand yielded by undercutting of the bluff is moved by the shore current and the beach-drifting process toward the point where it is constantly being built and rebuilt into beach ridges, spits, hooks, and looped bars by the waves and currents. Each change of wind having considerable duration probably results in a change of form of the structures. The hook at *B* is clearly older than the ridge at *C*. It was probably built by waves from the northeast or north-northeast which were somewhat refracted as they moved in along the shore. *C* was then built by larger waves which were perhaps from a more northerly direction, and such a large quantity of material was brought by the shore drift that it could no longer be carried westward to follow the outline of *B* but was deposited in the spit *C* to the south. As the water deepened and the waves were more refracted, so as to travel in a more westerly direction, the hook *A* was built. This was practically complete at the time of my visit, but there was a small outlet at *b* from the enclosed water to the lake. More recent southwest winds then caused the formation of the spit at *D*. All this was clearly shown in the relations of the various beach ridges.

Crystal Lake near Frankfort, Michigan, was lowered about 7 feet in the early seventies, exposing many shore structures and also starting new adjustments along the newly formed emergent shoreline. The lake is now surrounded by a broad sandy terrace which was for-

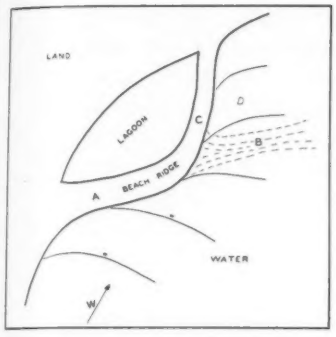

FIG. 6.—Looped bar, Silver Lake

FIG. 7.—Spit and hooked bar formation, Old Mission Point.

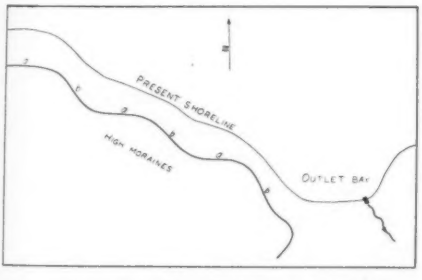

FIG. 8.—Portion of south shore of Crystal Lake

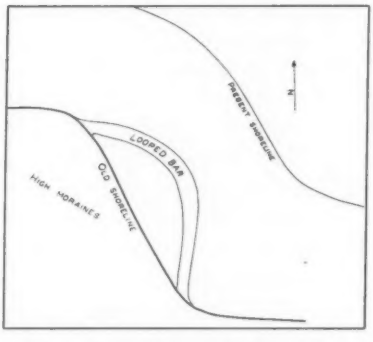

FIG. 9.—Looped bar, south shore of Crystal Lake.

merly lake bottom. On the south shore, west of Outlet Bay, the former bluffs of morainic material form a sawtooth arrangement of three projecting points (see Fig. 8). According to the United States Weather Bureau records, the strongest winds on this shore are from the west and northwest. Before the lowering of the lake the waves had cut deeply into the moraines, thus forming a terrace of glacial material extending out beneath the water 200 or 300 feet with a depth of 9 or 10 feet at the "dropoff." The material of this old terrace, the upper part of which is now above the water, consists of sand containing a considerable amount of gravel and large boulders. The sand and fine gravel were worked out to form part of the terrace, and some was carried southeast by the currents and shore drift. Some of this was laid down as looped bars in front of the old shore at *b*. These ridges, from 4 to 7 feet high and 20–40 feet wide, leave the old shores near the end of the promontories on their east side and gradually diverge from the cliffs until they are as much as 100 feet from them in some places. Then they swing back and again join the old shore. It is certain that one of them, which is shown roughly in Figure 9, formed a complete looped bar, but the other two may have been partly submerged at their southeast ends. These bars are, in general, steeper on the land side, showing that, after being deposited by the shore currents and before being brought above the water surface, they were worked in toward the shore by waves of translation, as so commonly happens with structures of this kind. No ridges are present on the old terrace in front of the cliffs at *a*, where the waves drove directly toward shore. The old terraces contain considerable sand which would be good material for the building of shore structures, but these features are found only along the cliffs at *b*, which lie nearly parallel to the direction of the prevailing winds. This is a strong indication that spits, bars, and related structures result from the combined work of the shore current and beach drifting rather than being thrown up by waves driving directly on the beach.

Since the lowering of the lake, considerable shifting of sediment has occurred in adjusting the shoreline to its new environment. This has resulted along this south shore in the building of large subaqueous dunes wherever the shoreline turns sharply toward the land. The

largest, Figure 10, is about 900 feet long and nearly half as wide. Along their lee sides to the southeast the depths increase abruptly from about 4 to 12 feet, with the sand generally lying at the angle of repose. According to De Beaumont's theory, these should be built up above the water by the waves, and since they are connected to the shore at one end they should become spits, but instead they are continuing to extend in length and to migrate shoreward beneath the water.

An examination of the shoreline shows the reason for the absence of spits. There is so much sand on the submerged part of the terrace that considerable quantities are brought into suspension by wave agi-

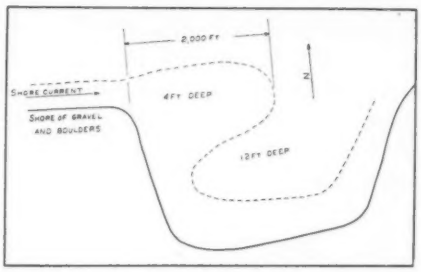

FIG. 10.—Sand deposit in Outlet Bay, Crystal Lake

tation and are transported readily by the shore currents. But at the new shoreline on the old terrace so much coarse debris occurs that very little beach drifting is possible. Because of this lack of beach material it is not possible for the submerged ridges to be built up above the water where they leave the shore, and as a result no spits form.

CONCLUSIONS

The above studies and observations indicate:

1. The subaqueous ridges which constitute the submerged portion of spits and bars are not built above a water surface of constant elevation by the work of the waves bringing up material from the bottom.

2. Spits increase in length above the water only when the waves and the shore drift move from the direction of their land connection. Waves from any other direction tend to destroy them.

3. Hooks are a normal accompaniment of spit-building, usually as a result of wave refraction. Neither change of wind direction nor the presence of tidal or hydraulic currents is necessary to their building.

4. The part of a spit above the water receives its material as the result of beach drifting. The direction the spit takes is determined by relations among the variables of current direction, wave direction, wave energy, amount of sediment available, and depth of water.

5. A looped bar is built by the same processes that produce spits and hooks. In fact, a looped bar is only a hooked spit that has continued to build until it has rejoined the shore.

Because the above summary disagrees somewhat radically with the conclusions of some previous investigators, it seems advisable to further simplify and explain some of the points. As far as my work shows, no ridges of fine sediment such as sand are brought by the work of the waves alone above a water surface having a constant elevation. However, it is possible, where rapid variations in water-level occur, as with tides, that a ridge formed at high tide by the plunging breakers might be exposed at low tide. A beach ridge might then be formed by the waves and be further elevated by dune-building; and so, under a fortuitous combination of conditions, an offshore bar might be built above the water. However, as far as I know, this process has never been observed.

Since offshore bars are not brought above the surface by wave work, it seems most logical to consider them as long spits that have been broken through by the waves. However, as stated by Johnson,[25] it is difficult to see how such great quantities of material could be transported along the shore the necessary distances for the building of such structures. It is possible that some of the material in offshore bars may be brought from the bottom and added to the spit by waves in the same way they sometimes add to a beach ridge from a shallow terrace. This sand would then be subject to beach drifting and thus be added to the spit along with that brought directly from the land. However, the whole subject of offshore bars needs further study.

The statement that hooks and looped bars are usually built by

[25] *Ibid.*, p. 355.

waves coming from the same direction as those that built the spit does not mean that tidal or hydraulic currents across the end of the spit might not result in hook formation. Also, an onshore wind blowing perpendicular to the axis of a spit causes beach drifting across the end of it and so aids in the formation of the hook. However, refracted waves are usually the chief agent in the process.

Johnson[26] states that bay bars are sometimes formed and brought above the water by the breakers. While, according to my observations, these ridges are never brought above the water surface by direct wave work, the foundation of the bar in the form of a subaqueous ridge might be formed by the plunging breakers as described in the building of the low and ball, and this might be followed by the actual building of the ridge above water by the processes of shore drifting from one or both sides.

[26] *Ibid.*, p. 301.

THE ASOTIN STAGE OF THE SNAKE RIVER CANYON NEAR LEWISTON, IDAHO[1]

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ABSTRACT

An early chapter in the evolution of the Snake River is recorded by intracanyon lavas in and near the Lewiston Basin of Washington and Idaho. Dissection, attendant upon the beginning of late Cenozoic deformation of the Columbia River lavas, produced canyons which reached a maximum known depth of more than 1,200 feet before they were nearly filled with two thick lava flows. This stage of dissection, which is termed the "Asotin" stage, and the volcanism which terminated it are probably of early Pleistocene age.

The Snake River canyon of the Asotin stage was 900 feet deep south of the axis of the Lewiston downwarp and more than 1,200 feet deep on the north, or downstream, side of the axis, thus suggesting an early stage in the folding of the Lewiston downwarp. Continued regional deformation has brought the bottom of Asotin Canyon about 300 feet below present river-level at the axis of the downwarp and elevated it nearly 1,000 feet above river-level on the margins of the downwarp.

The Snake River, during post-Asotin time, followed on and near its former course and carved a new canyon that is deeper than the Asotin Canyon except in the center of the Lewiston downwarp. Much of the intracanyon lava was removed during the later stage of canyon-cutting, but a number of remnants remain in a known distance of 22 miles.

INTRODUCTION

Near the southeast corner of Washington and north of the Wallowa–Seven Devils uplift, the Snake River enters an extensive area of Tertiary lavas in the Columbia Basin. The river flows northward along the Idaho-Washington boundary in a canyon of decreasing depth to the Lewiston Basin. There, augmented by the Clearwater River (Fig. 1), the Snake River turns abruptly westward and follows a westerly course for 8 miles across the basin; then, turning northward again, it leaves the basin through another deep, narrow canyon.

The even continuity of the widespread Columbia River lavas is well displayed in the canyon walls. These layered flows are interrupted in a number of places along the canyon near Lewiston, Idaho,

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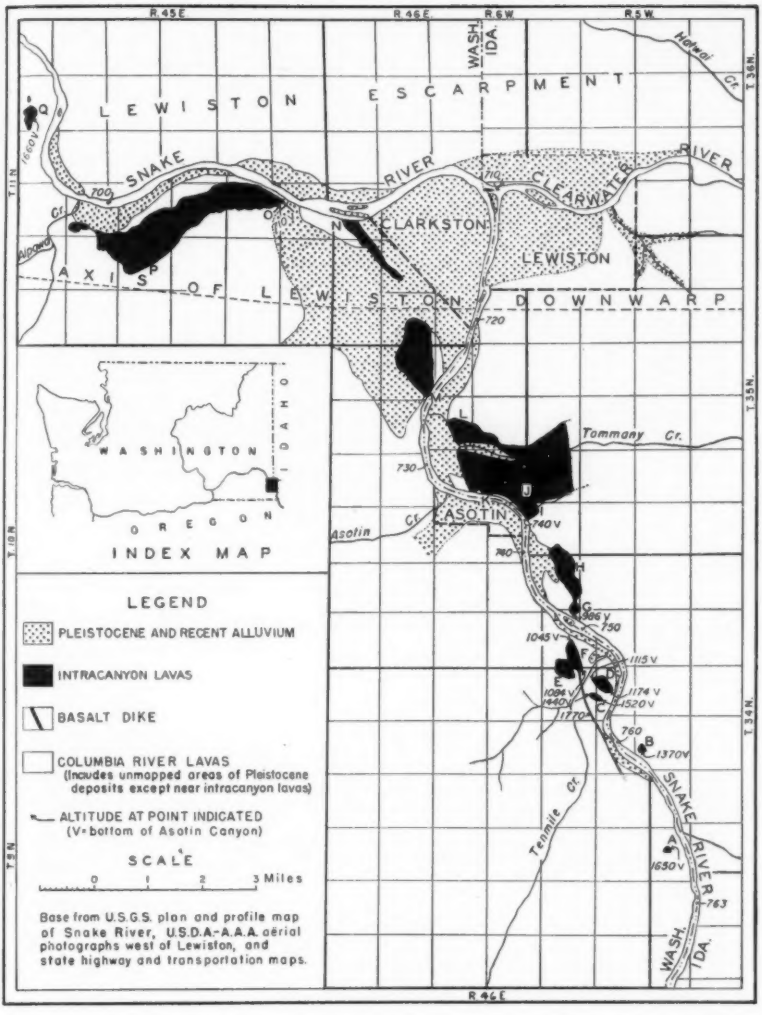

FIG. 1.—Geologic map of area adjacent to Snake River in the vicinity of Lewiston, Idaho, showing distribution of remnants of lava that filled Asotin Canyon. Lava remnants are lettered for reference in the text.

FIG. 2.—Cross section of the lower part of Asotin Canyon and lavas on northeast side of Snake River near Tenmile Creek (see *G*, Fig. 1). The sheer rock face in center of view is columnar basalt of the lower flow in Asotin Canyon. Columns radiate from center toward canyon sides. Grass-covered slope above cliff and skyline knob are lower part of upper flow.

FIG. 3.—Oblique view of intracanyon lavas 2 miles west of Clarkston (see *O*, Fig. 1), as seen from the north side of Snake River. The intracanyon lavas form the core and backbone of the ridge. Columbia River lavas, showing in ledges below the contact, lie in a narrow blade between Asotin Canyon and the modern canyon.

by V-shaped and semicircular patches of lavas and related fragmental materials (Figs. 2, 3) which extend across flows to a maximum observed depth of 850 feet. Some of these patches were mapped by R. E. Fuller and interpreted as cross sections of volcanic vents of an unusual type. He named them the "Asotin craters" for the town of Asotin, Washington. The following explanation of their origin is quoted from the abstract of his paper: "These [vents] furnish proof of explosive origin. Evidence is offered to prove that this activity has been due to the contact at depth of basaltic fissure eruptions with water-saturated gravels."[2]

The resemblance of the Asotin craters to erosional remnants and cross sections of a former lava-filled canyon prompted the present writers to restudy these features, and the study showed that the lava of these craters had flowed into and locally filled an old river-cut canyon with typical gradient, profiles, and other erosional features.

These patches of lava give clear-cut evidence of an early stage in the history of the Snake River during which a canyon was cut to a maximum known depth of more than 1,200 feet before it was largely filled with two thick lava flows. The lava flows subsequently were folded and finally were dissected by later erosion. The recognition of this episode in the geologic history of the region provides information bearing on various physiographic and stratigraphic problems of the late Cenozoic.

REGIONAL GEOLOGY

The geology of the Lewiston region has been well described by I. C. Russell,[3] and only a brief summary description will be given here. The whole area is underlain by a series of basaltic lavas and minor interbedded sedimentary units and is situated within the eastern margin of the widespread Columbia River lava of Russell.

A close correlation exists between the present topography and the geologic structure. The lavas are nearly horizontal on the plateau north of Lewiston but rise slightly southward to the edge of the Lewiston escarpment. There, along an east-west axis, the lavas bend

[2] "The Asotin Craters of the Columbia River Basalt," *Jour. Geol.*, Vol. XXXVI (1928), pp. 56–74.

[3] "Geology and Water Resources of Nez Perce County, Idaho," *U.S. Geol. Surv. Water-Supply Paper 53* (1901).

down abruptly at angles of as much as 60°, and their top surface descends from an altitude of more than 2,700 feet to an altitude of less than 1,000 feet at the synclinal axis of the downwarp near Lewiston and Clarkston. Southward from the axis of the downwarp the surface of the lavas rises at an average rate of less than 3° to the margin of the area studied. The canyon of the Snake River has been cut about 300 feet below the top of the lavas in the trough of the syncline to an altitude of less than 700 feet above sea-level. Thus, the Lewiston Basin is the topographic expression of an elongate asymmetric downwarp with gentle southeast and southwest slopes and a steep north slope. The abrupt escarpment on the north side of the basin is 2,000 feet high and is essentially a dip slope only slightly modified by erosion.

THE ASOTIN STAGE

GENERAL STATEMENT

It is generally agreed that the upper surface of the Columbia River basalt was essentially a plane surface at the end of the period of extrusion of these lavas and that subsequent crustal movements appear to have been negligible until late in the Tertiary or until the beginning of the Quaternary.

Throughout the latest Miocene and the Pliocene, eastern Washington and eastern Oregon seem to have been characterized by reduction of the surrounding and protruding highlands of pre-Tertiary rocks and spreading of sediments and lavas in the lowlands. Abruptly this cycle of planation changed to one of deep canyon-cutting. The Snake River canyon is 2,000 feet deep where it crosses the folded lavas in the Lewiston escarpment, but it decreases in depth toward the center of the Columbia Basin until there is no canyon. Regional deformation appears to have accompanied or preceded the development of this canyon. Similar evidence of rejuvenated erosion and deformation is present in many places in the Columbia Basin. However, the present canyon of the Snake River in the Lewiston area is not the original canyon of that stream in that area, as an early canyon was cut to a depth of more than 1,200 feet and was then nearly filled with lava before the present canyon was cut. No canyon of comparable size is known from the central part of the Columbia

Basin, and it may be presumed that the original Snake River canyon near Lewiston was initiated by the same regional deformation that caused the development of the present canyon. This early period of erosion, which was terminated by volcanism, is here called the "Asotin" stage of the Snake River.

The Snake River now follows closely the lava-filled canyon of the Asotin stage, crossing and recrossing it. As a consequence, more than half of the intracanyon lavas were removed during the development of the present canyon. Their remnants are preserved as patches clinging to the canyon walls, as discontinuous bodies cutting across spurs of the canyon walls, and as more extensive remnants as much as $3\frac{1}{2}$ miles in length.

ASOTIN CANYON

Profile.—The lava-filled Asotin Canyon as seen in cross sections exposed in the canyon of Snake River and tributary gulches commonly appears as broad, V-shaped, roughly semicircular, or parabolic sections of fresh-appearing volcanic materials that interrupt the even continuity of the older lavas in which the canyon was cut. The fill is locally more resistant to erosion than the older lavas, and its course is commonly marked by a prominent rocky ridge.

The Asotin Canyon was similar in profile to modern canyons of comparable depth in the region, though it was somewhat narrower and more steeply walled. The cliffs and ledges on the sides of the Asotin Canyon show the same close relation to the individual flows of the Columbia River basalt as the cliffs and ledges along the present canyon. An arkosic sandstone in the upper part of the Columbia River basalt section has had a marked effect on the topography throughout the Lewiston Basin. The sandstone is nearly 250 feet thick near Asotin and is overlain by lava flows from 100 to 300 feet thick. Erosion of the weak sandstone has produced either gentle slopes or prominent terraces that are surmounted by a rimrock of lava. These same features were present along the Asotin Canyon; its normal profile possessed a narrow inner canyon ranging from 300 to 600 feet in depth below the interbasalt sedimentary member and a more widely flaring upper portion ranging from 200 to 400 feet in depth and surmounted by steep rims.

The twofold character of the Asotin Canyon is responsible for some features of the distribution of the intracanyon lava that appear incongruous on the map. The width of outcrop of the lava is more than 1 mile along lower Tammany Creek, where a tributary entered the Asotin Canyon from the east, whereas along Snake River above Asotin the width of the outcrop is only 2,000 feet. However, the lava remnants along that part of the Snake River represent the lava filling of the narrow inner canyon, whereas the lava along the lower part of Tammany Creek represents the lava that spread more widely in the flaring upper part of the canyon over the stripped rock bench at the base of the arkosic sandstone. Possibly the canyon was actually wider at the mouth of the tributary than elsewhere.

The lava remnant on the west side of Tenmile Creek is unusually wide, because the flow in the main canyon (Fig. 1, *D* and *F*) crossed a low divide into a nearly parallel tributary canyon (*C*, *E*).

Depths and present altitudes.—As pointed out above, the present canyon is usually surmounted on each side by rimrock. As will be shown later, the Asotin Canyon probably postdates the rimrock flows, so that depths of that canyon can be measured from the rimrock.

The depth of Asotin Canyon from the most southerly known fill remnant at *A* (Fig. 1) northward to the axis of the Lewiston downwarp was between 850 and 900 feet. The northernmost known remnant at *Q* lies high on the west side of the Snake River Canyon north of the point where the river leaves the downwarp through the 2,000-foot escarpment of steeply inclined basalt flows. Here the Asotin Canyon was cut to a depth of at least 1,225 feet, or more than 325 feet deeper than on the south side of the downwarp. The greater depth is perhaps explainable by an abrupt and marked increase of gradient fortuitously located on the axis of Lewiston downwarp, but the increase in depth is more likely a measure of the amount of structural closure developed in the Lewiston downwarp by the end of the Asotin stage.

Post-Asotin deformation has depressed the base of the fill well below river-level at the axis of the Lewiston downwarp. South of the axis the bed of the Asotin Canyon rises to the level of Snake River directly east of Asotin and continues to rise southward at the rate of

130 feet per mile. This gradient projected northward to the axis of the Lewiston downwarp would give a depth of about 500 feet below river-level for the bed of Asotin Canyon at the axis. However, the gradient undoubtedly lessens near the axis, and estimates based on the thickness of the lava fill suggest that the depth below the present river-level should be of the order of 300–350 feet. The dip of the lowest exposed flows of the Columbia River basalt in the canyon south of Asotin is about 170 feet to the mile, or 40 feet to the mile greater than the gradient of the floor of the Asotin Canyon. However, the flows of Columbia River basalt above the sedimentary zone near the top of the canyon have approximately the same dip as the gradient of the Asotin Canyon, and there is no appreciable variation in depth of the Asotin Canyon in this part of the area. The bed of Asotin Canyon north of the axis of the Lewiston downwarp rises to river-level about 1 mile west of Clarkston and then rises westward only about 50 feet in 4 miles to Alpowa Creek in a direction nearly parallel to the axis; north of Alpowa Creek the bed of the canyon rises rapidly northward to an altitude of 965 feet above the river at *Q*.

Walls and floor.—Gravel is present on the floor of the Asotin Canyon at all localities where the floor is well exposed. The pebbles and boulders are well rounded and consist largely of metamorphic and plutonic rocks of the same type that were transported in large quantities by the Snake River during the Pleistocene and Recent epochs. The gravel matrix is micaceous arkosic sand. The gravels are restricted to the bottom of the old canyon and range from 1 foot to more than 30 feet in thickness. They lie upon a fresh stream-worn rock surface; and, in many places, as on the east side of Tenmile Canyon and along the Snake River opposite Asotin, they are packed in potholes. The old canyon bottom also shows much basalt debris and large rolled blocks derived from the sides of the Asotin Canyon. These blocks commonly show a deep rind of spheroidal weathering which contrasts sharply with the fresh surface of the overlying intracanyon lavas or the occasional pillow or bomb in the palagonite which commonly separates the stream gravels from the intracanyon lavas.

The walls of the Asotin Canyon show fresh rock surfaces of Columbia River basalt in a few places but more commonly are considerably

weathered. Soil layers as much as 3 feet deep are also common. The soil is composed largely of decomposed basalt bedrock and detritus, as well as slope wash from the interbasalt arkosic sandstone which lies near the top of the Columbia River basalt section. The Columbia River basalts show much reddish and brownish iron-oxide stain, which contrasts with the gray and black of the noticeably fresher intracanyon lavas.

INTRACANYON LAVAS

General statement.—The largest remnants and the deepest sections of the intracanyon lavas are preserved in and near the middle of the Lewiston downwarp, where erosion by the present Snake River has been least. Only small, shallow remnants of the intracanyon lavas, such as those at *A*, *B*, and *Q*, remain on the margins of the downwarp where the present canyon has been cut far below the bottom of Asotin Canyon. The fill remnants at *A*, *B*, and *Q* each preserve less than 400 feet of the original depth; but east of Asotin at *J* and west of Clarkston at *P* the eroded surfaces of the intracanyon lavas still rise to the level of the topmost flow of Columbia River basalt, showing a minimum depth of 850 feet for the lava fill. In this part of the basin the fill is composed of two thick basalt flows. Northeast of Asotin, where nearly the full depth of the fill is preserved, the lower flow is 330 feet thick and the upper flow is at least 520 feet thick.

The lower flow.—The lower flow was confined to the narrow inner gorge of Asotin Canyon. It is present in all the known remnants of the intracanyon lavas and forms imposing rock faces and sheer cliffs (Figs. 2, 3) where it is exposed in the walls of the Snake River canyon. Both hackly and columnar basalt are present, and the columnar structure is strikingly developed in nearly every exposure. The columns are exceptionally slender, and in some places columns about 1 foot in diameter reach a length of 75 feet. They are most pronounced near the contact of the flow with the walls of the old canyon, less pronounced in the central part of the flow, and generally absent at the top. The orientation of the columns was controlled largely by the attitude of the cooling surface at the canyon sides, and the columns are perpendicular to that surface. This produces giant fanlike rosettes in the narrow canyon bottoms. Much of the central part of

the lower flow is solid basalt with closely spaced hackly fracture. The upper part shows vesicular, scoriaceous, and fragmental materials, in places to a depth of more than 50 feet. A layer of palagonite, nearly everywhere less than 6 feet thick, is present in most places at the base of the lower flow. Differentiation of the flow into minor sheets has been observed only in the cliffs opposite Asotin and west of Clarkston at *N*, where they appear to be marginal facies of small extent.

FIG. 4.—View northwest down lower Tammany Creek. Hill on right, Swallow's Nest cliff in middle distance (*M* of Fig. 1), and rounded hill beyond Swallow's Nest are remnants of the massive solid portion of the upper flow of intracanyon lavas. Pleistocene silts and gravels on left; Lewiston escarpment in distance.

The upper flow.—The upper flow occurs mainly near the axis of the Lewiston downwarp, and thick sections of the upper flow have been found only in that part of the area. Elsewhere it is largely or entirely eroded away.

In the high hills between Tammany Creek and the Snake River, as well as in the valley of Tammany Creek, basalt with hackly fracture is 500 feet thick and shows no marked divisions into minor flows. This massive flow continues northwestward across the Snake River to the great cliff that is known as the "Swallow's Nest" (Fig. 4, *M*). At the Swallow's Nest and in the hill immediately to the northwest a thickness of 475 feet of the upper lava flow is exposed with only the

upper 25 feet of the lower flow above water-level at the base of the cliff. No division of the upper lava into sheet flows is apparent in the main cliff, but on the south side of the Swallow's Nest the lower part of the massive flow grades into minor sheets which intertongue with palagonite breccias known to local stonemasons as the "Asotin sandstone." Similar features are found elsewhere, indicating that the upper part of the intracanyon lava is a single flow with marginal facies of thin sheet flows and fragmental products.

The marginal facies of this flow are more commonly exposed than the more massive central facies and also provide more interesting features. Some of the sheets are vesicular and craggy; others are largely solid with columnar or hackly structure. Some sheets are horizontal, and others are inclined at angles as high as 35°. Palagonite and breccia occur as irregular and lenticular masses ranging from a few feet to more than 200 feet in thickness. These masses are principally concentrated along the sides of the old canyon at and near the base of the upper flow, though the breccia is also found higher up on the sides as a distinct contact phase several feet thick and as irregular masses crowded onto ledges and into re-entrants by the moving lava.

One of the best places to study the upper flow is in the cliffs across the Snake River from Asotin at *K*. Here the lower flow forms a nearly sheer cliff that rises 240 feet above river-level; above this, at the base of the upper flow, lies a mass of palagonite about 200 feet thick in which are some thin lava sheets and coarse breccias. Along the cliffs and bluffs that extend eastward along the river and up a small tributary canyon, the palagonite grades into a series of thin, nearly horizontal lava sheets with an aggregate thickness of about 400 feet. A layer of palagonite from 5 to 20 feet thick underlies the series of lava sheets.

Source of the intracanyon lavas.—No reliable information on either the source or the direction of flow of the intracanyon lavas has been obtained. Probably the lava flowed down the river, since the general lack of fragmental products at the base of the lower flow suggests that the lava did not encounter a large amount of water in its course. No evidence was found of a lava flow entering the canyon from the plateau surface, though this may have taken place beyond the region in which remnants of the canyon fill are known. A dike is exposed

in the canyon of lower Tenmile Creek and in the canyon of the Snake River to the southeast. It intersects the main Asotin Canyon beneath talus on the west side of Tenmile Creek, and at one time it intersected a tributary of the Asotin Canyon in at least two places farther south. The dike is nearly vertical and ranges from 5 to 20 feet in width. It is composed of unusual materials, for it abounds in vesicular basalt and in breccia and other fragmental materials. The more solid portions are commonly made up of a series of nearly vertical lenses or layers with vesicular borders and here and there some rude columnar jointing. The dike, therefore, does not contain materials to be expected in a feeder of either the massive intracanyon lavas or the Columbia River basalts.

LATER EROSIONAL HISTORY

The Asotin stage closed with the filling of the old canyon by the intracanyon lavas, but it may be desirable to trace the subsequent erosional history because it affords some evidence on the age of the Asotin stage.

The Snake River was not completely expelled from its old course by the intracanyon lavas but continued to flow along one margin or the other of the lava fill throughout the early Pleistocene, crossing back and forth across the lava fill at least six times within the known extent of Asotin Canyon. The failure of the intracanyon lavas to expel the river permanently from its old course, even though the old canyon was locally filled to overflowing with the lavas, may be due to the development of a shallow channel on the surface of the intracanyon lava as a result of shrinkage as the thick flows cooled or of subsidence of the top of the flow as the lava of the core flowed on downstream. The latter hypothesis is suggested by many structural features of these intracanyon flows. Erosion of the canyon fill was accompanied by deformation, and the two were closely interrelated. On the margin of the Lewiston downwarp the floor of the old canyon was raised, and a new canyon was cut down through the intracanyon lavas and far below their base. At *Q*, where the river cuts through the Lewiston escarpment, the floor of the old canyon was elevated 965 feet above present river-level, and the present canyon is thus nearly 1,000 feet deeper than the old one. On the other hand, the floor of the old canyon is estimated to lie more than 300 feet below

river-level near the axis of the Lewiston downwarp, and the modern canyon is shallower than the old one. Despite these marked effects of deformation the Snake River continued to occupy its former channel to such an extent that, with one exception, the floor of the Asotin Canyon is nowhere more than a mile from the course of the present river. The only part of the present course of Snake River that diverges widely from Asotin Canyon is north of Asotin, where the Snake River flows northward to Lewiston in a late Pleistocene diversion channel. The post-Asotin channel from which the river was diverted in the late Pleistocene followed closely along the southwest side of the Asotin Canyon from Asotin northwestward and received the Clearwater River 4 miles west of its present mouth. This abandoned canyon of post-Asotin age is now filled with cemented, iron-stained, glacial gravels to a depth of 400 feet. Thus the post-Asotin erosional history of the Snake River includes erosion of a new channel essentially along its former course in the early Pleistocene and, locally, the erosion of a third channel in the later Pleistocene.

AGE OF THE ASOTIN STAGE

The intracanyon lavas are clearly younger than the Columbia River basalt exposed in the walls of the Snake River Canyon, for east of Asotin the intracanyon lavas lie against the eroded edges of the Columbia River basalt within 50 feet of its upper surface. The intracanyon lavas are noticeably less weathered than the older lavas. Furthermore, if the intracanyon lavas had been buried beneath recurrent Columbia River basalt, the present course of the Snake River would probably have been diverted from its earlier course in Asotin Canyon.

The main mass of the Columbia River basalt is generally considered to be Miocene. The Asotin stage is presumably younger than Miocene and may be compared in age with some of the post-basalt sedimentary formations.

The Ringold formation, containing late Pliocene or early Pleistocene vertebrate fossils,[4] overlies the Columbia River lavas in central Washington. The relation of the Ringold formation to the deforma-

[4] J. C. Merriam and J. P. Buwalda, "Age of Strata Referred to the Ellensburg Formation in the White Bluffs of the Columbia River," *Univ. Calif., Dept. Geol. Sci. Bull.*, Vol. X (1917), pp. 255–66.

tion of the Columbia River lavas is important for the purpose of dating the Asotin stage during which deformation began. Merriam and Buwalda favored the belief that the Ringold was deposited in structural and erosional depressions after the deformation, but they called attention to the tenuous nature of the evidence then at hand. Harold E. Culver[5] contends that the sediments of the Ringold were spread widely over the Columbia River basalt prior to deformation. His conclusion is based in part upon the doubtful evidence of lithologic similarities between folded and faulted post-basalt sediments, such as those on the Saddle Mountain uplift, with the Ringold formation at the type locality. Nevertheless, his correlation of sediments west of Washtucna with the Ringold formation is borne out by the finding of a typical Ringold fauna at McChesney Springs west of Washtucna. There the Ringold formation lies with apparent conformity upon the basalt at an altitude about 1,000 feet above that of the base of the formation in the type locality. Thus there can be no doubt that the Ringold is present in tracts elevated during the deformation and that it has been dissected during the present erosion cycle. C. R. Warren[6] concludes that deposition of the Ringold formation was contemporaneous with the deformation, although his conclusion is based upon a postulated physiographic history in which are various uncertainties. The general absence of basaltic gravels in the Ringold formation is possibly the best evidence that the Ringold is not younger than the deformation. Since evidence has been presented that the Asotin stage was initiated by the deformation, it is probable that the Asotin stage is younger than the Ringold formation. The youthful stage of erosion in the Lewiston region also suggests that the area has been subjected to active erosion for only a relatively short time.

An upper time limit for the Asotin stage is established by the indurated gravels, sands, and clays which partly filled[7] the post-Asotin canyons of the Lewiston region. These deposits include blue clays and ice-rafted erratics, which are indicative of Pleistocene age; but

[5] "Extensions of the Ringold Formation," *Northwest Science*, Vol. XI (1937), pp. 57–60.

[6] "Course of Columbia River in Southern Central Washington," *Amer. Jour. Sci.*, Vol. CCXXXIX (1941), pp. 209–32.

[7] R. L. Lupher, "The Clarkson Stage of the Snake River," manuscript in preparation.

their induration and degree of weathering indicate that they are much older than the scabland deposits and Touchet beds of the later Pleistocene.[8] The available evidence thus appears to indicate that the Asotin stage is probably of early Pleistocene age.

PHYSIOGRAPHIC AND STRATIGRAPHIC IMPLICATIONS OF THE ASOTIN STAGE

The Asotin stage marks the beginning of the late Cenozoic diastrophism and erosion which produced the main physiographic features of the Columbia River lava region. The effects of the early erosion were largely nullified in the Lewiston region by the extrusion of lavas into Asotin Canyon, forcing the Snake River to cut new canyons which followed approximately their former courses.

Lava flows occupying canyons and valleys formed after the extrusion of Columbia River lava have been reported by a number of writers in the Cascade region, in central and southeastern Oregon, and in southern Idaho, but these lavas range in age from Pliocene to Recent and are in regions of long-continued volcanic activity. Such features are less evident in the northern and eastern regions of the Columbia River lava region, although J. T. Pardee and Kirk Bryan[9] have reported lavas in valleys eroded in the topmost Columbia River lava flows and the underlying Latah formation of the Spokane region. The valley flows of the Spokane region may well be of the same age as the intracanyon lavas of the Lewiston region. Correlation of the Asotin Canyon with folding of the Columbia River basalts emphasizes the possibility that basaltic extrusions were renewed in adjacent areas after the beginning of the structural ridges and basins of the lava region.

Additional study of the Asotin Canyon will undoubtedly furnish some critical evidence on the history of the Snake River. One implication is already evident in the size of the stream that made the canyon. It was a small stream, in comparison to the present Snake River, perhaps less than 100 feet wide in some places. It carried gravels that are much finer than the coarse cobble gravels of the

[8] R. F. Flint, "Origin of the Cheney-Palouse Scabland Tract, Washington," *Bull. Geol. Soc. Amer.*, Vol. XLIX (1938), pp. 461–524.

[9] "Geology of the Latah Formation in Relation to the Lavas of the Columbia Plateau near Spokane, Wash.," *U.S. Geol. Surv. Prof. Paper 140* (1925), pp. 1–16.

modern Snake. Possibly it did not consist of the great, fully integrated system that we know today. Perhaps the Snake River above the Seven Devils country belonged to some other system, and the Asotin Canyon carried mainly drainage from the Grande Ronde and Salmon River systems. The abundant metamorphic and plutonic rocks in the gravels of Asotin Canyon are like those now coming from the Salmon River country.

More complete knowledge of the Asotin Canyon should also furnish some information on a peculiar feature of the entrenched meanders along the Grande Ronde River. The Grande Ronde, rising near La Grande, Oregon, flows northeast for 90 miles through entrenched meanders which reach a depth of 3,000 feet along its lower course. It joins the Snake River only 11 miles south of the southernmost known remnant of the intracanyon lavas, and here the meander pattern abruptly ceases. The detailed course of the bed of Asotin Canyon cannot be traced between the various cross sections where the canyon bottom is exposed. Nevertheless, the orientation of fill remnants and the asymmetric canyon profiles indicate some pronounced bends in that stream east of Asotin and east of lower Tenmile Creek. These are not comparable to some of the striking loops of the modern entrenched Grande Ronde but may compare with the Grande Ronde meanders in their early stage of entrenchment. Future investigations may show whether the intracanyon lavas were instrumental either in forming the meanders above the mouth of the Grande Ronde or in destroying them below that point.

Another implication is that the Clearwater River formerly had a different course. No remnants of intracanyon lava have been found in the present canyon of the Clearwater River. It is possible that, at the time of extrusion of the intracanyon lavas, the Clearwater canyon was so small that the lava could not flow any distance up the canyon or that, if it did, the remnants have all been cleaned out. Another possibility is that the Clearwater River did not then occupy its present course. The large tributary branch of the Asotin Canyon along lower Tammany Creek, 5 miles south of the present Clearwater River, seems disproportionately large for a creek such as Tammany Creek and may well represent the site of the ancestral Clearwater River.

CAMBRIAN SEQUENCE IN WESTERN GRAND CANYON, ARIZONA[1]

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ABSTRACT

Correlation of Cambrian strata in the western part of the Grand Canyon involves consideration of formations and faunas from eastern and southeastern Nevada and from the Granite Gorge area of the Grand Canyon in Arizona. Lithologic units defined in both regions are recognized. With priority of definition as the basis for acceptance of formational names, all units to which such definitions apply have their type localities in the Basin Range region.

The Cambrian sequence near Grand Wash Cliffs in the westernmost part of the Grand Canyon contains 179 feet of Lower Cambrian Prospect Mountain quartzite (Eureka district, Nevada), which rests unconformably upon pre-Cambrian granite gneiss. These rocks are overlain by 304 feet of Lower and Middle Cambrian Pioche shale (Pioche district, Nevada), carrying the *Olenellus* zone fauna at the base and *Albertella* near the top. Above the Pioche formation lie 70 feet of strata consisting of two massive dolomite members separated by 40 feet of sandstone, shale, and dolomite. This unit is the lithogenetic equivalent of the Lyndon limestone (Pioche district). The Lyndon is followed by 99 feet of Chisholm shale (Pioche) with the *Glossopleura* zone fauna. The Chisholm formation is in turn overlain by about 380–400 feet of Peasley limestone (Pioche). All the above-mentioned Cambrian formations are conformable in western Grand Canyon. The last three are of medial Cambrian age.

Disconformably above the Peasley limestone lie 110 feet or less of limestones of probable medial Cambrian age. These beds, which here are named the "Mead formation," are disconformably overlain by a thick series of undifferentiated dolomites and limestones of Cambrian(?) and post-Cambrian Paleozoic age.

INTRODUCTION

The Cambrian section exposed in western Grand Canyon near Grand Wash Cliffs at the western edge of the Colorado plateau in northern Arizona is critically situated (Fig. 1). The value of this section is threefold: first, these Cambrian strata are more abundantly fossiliferous than those of most areas in the Southwest; second, the location of this section on the boundary between the Colorado Plateau and Basin Range provinces provides a long-needed tie between the Cambrian deposits of these two physiographic regions;

[1] Published by permission of the director, National Park Service, and the director, Nevada State Bureau of Mines.

and, third, this tie is facilitated by the fact that during much of Cambrian time the sequence of events in the area near Grand Wash Cliffs was almost synchronous with that of the classic Pioche, Nevada, region to the north.

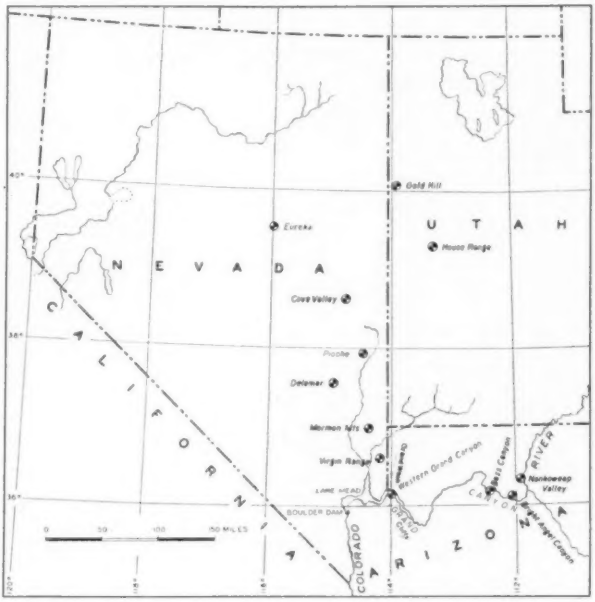

FIG. 1.—Index map of a portion of southwestern United States, showing location of Cambrian occurrences mentioned in this report.

The associate writer[2] is engaged in the preparation of a series of Lower and Middle Cambrian formational and faunal correlations in eastern and southeastern Nevada and northern Arizona. The western Grand Canyon sequence is one of this series. However, because of the importance of this Cambrian section, together with the cooperative character of the investigation, this stratigraphic account is presented as a separate contribution.

[2] Wheeler, "Lower and Middle Cambrian Correlations in Nevada and Arizona" (abstr.), *Bull. Geol. Soc. Amer.*, Vol. LII (1941), pp. 1959–60.

Field work was begun by Schenk in June, 1936, in connection with general studies of that portion of the Boulder Dam National Recreational Area subsequently inundated by the waters of Lake Mead. The writers visited the area in December, 1938, in company with Dr. Dwight M. Lemmon. Schenk and Edwin D. McKee were in the

FIG. 2.—South wall of western Grand Canyon near Grand Wash Cliffs, Arizona. The foreground block is downdropped on the opposite shore. *Cps* = Pioche shale; *Cl* = Lyndon formation; *Cc* = Chisholm shale; *Cpl* = Peasley limestone; *Cm* = Mead limestone.

field in February, 1937, and April, 1939. During early November, 1940, the writers completed the field work in company with Dudley L. Davis.

Various Cambrian sequences have been examined in the Lake Mead–Grand Wash Cliffs region. The section chosen as typical of this area and which is described in this report is located on the south wall of the Grand Canyon, about $1\frac{1}{2}$ miles east of Grand Wash Cliffs and about $\frac{1}{2}$ mile west of Rampart Cave (Fig. 2).

NOMENCLATURE

Regional and interregional correlation of formations usually entails certain difficult problems of stratigraphic nomenclature. The degree of difficulty encountered depends upon both the depositional facies relationships and the many factors which may affect changes subsequent to sedimentation. Except for certain phase conditions, such as secondary dolomitization of limestones, most of the more widespread Cambrian units of the Southwest are lithologically recognizable from place to place. The principal problems, therefore, are those commonly encountered in instances of marine oscillations involving overlap and offlap near a geosynclinal margin.

As stated above, the western Grand Canyon Cambrian has been studied in the light of its relationships to strata both in the Basin Range and in Colorado Plateau provinces. Wherever possible, correlations have been made with the type localities of previously designated formations in neighboring regions. In every case, priority of definition is employed as the basis of selection of formational names.

In this regard the reader may question the acceptance of five Nevada formational names, some of which have been proposed only recently, whereas none is brought in from the Colorado Plateau. Those who are familiar with the western Grand Canyon Cambrian sequence are impressed with its similarity to that found at Bass Canyon and vicinity, nearly 100 miles to the east. For example, G. K. Gilbert[3] long ago referred some of these strata to the Tonto group; and the presence of the lithogenetic counterparts of the "Tapeats," Bright Angel, and Muav formations near Pierce Ferry is cited by R. B. Wheeler and A. R. Kerr.[4] The name "Tapeats" is rejected here on the basis of priority. The strata equivalent to the Bright Angel shale in the Pioche, Nevada, district, have been subdivided successfully into three formations: Pioche, Lyndon, and Chisholm. These three formations are lithologically, stratigraphically, and faunally recognizable at Delamar, Mormon Range, Virgin Range, and elsewhere between Pioche and the western terminus of the Grand

[3] *Geog. and Geol. Explorations and Surveys W. 100th Merid.*, Vol. III, Part I (1875), p. 163.

[4] "Preliminary Report on the Tonto Group of the Grand Canyon, Arizona," *Contr. Grand Can. Geol., Nat. Hist. Bull. No. 5* (1936), pp. 6–10.

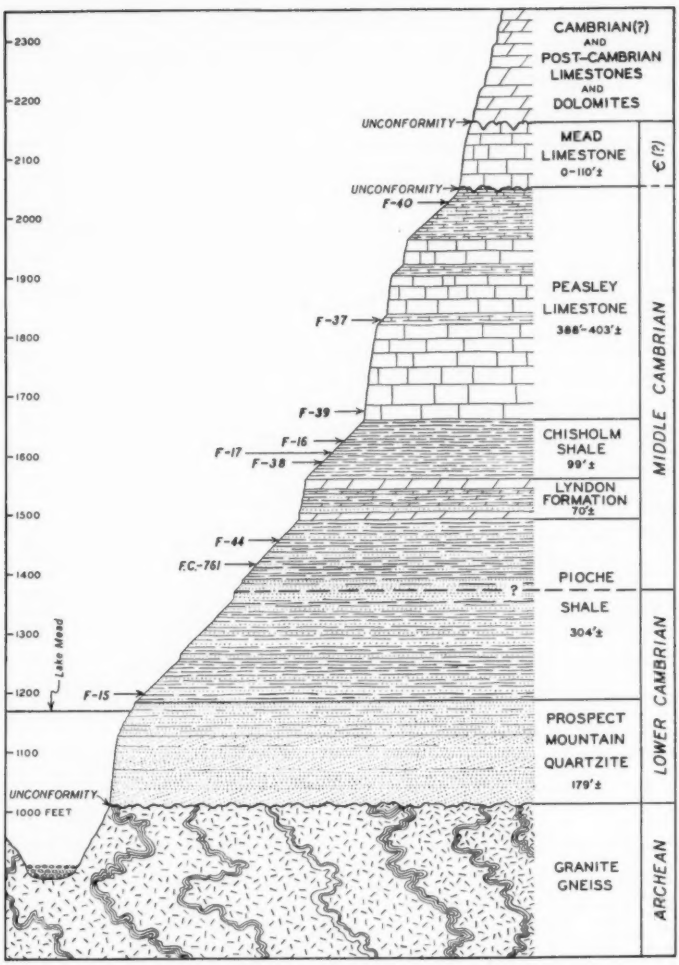

FIG. 3.—Columnar section, with horizontally shortened profile, showing relationships of Cambrian formations on the south wall of western Grand Canyon near Grand Wash Cliffs, Arizona.

Canyon. These units may be regarded, therefore, as traceable. Thus, in the last analysis, priority of definition is applied again here. Although until recently we had intended to employ the name "Muav" limestone,[5] this term is now abandoned in the Lake Mead region for reasons to be discussed on a later page. In only one instance, that of the Mead formation, is there sufficient doubt of formational identity to warrant the application of a new name.

That these formations have their lateral limits of recognition is appreciated. This becomes especially apparent as a comparison is made between the rocks of the section herein described and those farther east in the Grand Canyon. As the eastern margin of the Cordilleran geosyncline is approached, the Bright Angel shale, though somewhat younger, is the lithogenetic equivalent of the combined Pioche, Lyndon, and Chisholm formations. The Bright Angel shale at its type locality, however, is not subdivisible into these same three formations. Therefore, the Bright Angel remains a valid unit toward the eastern margin of the trough. Additional examples of this kind could be cited; and, as the southern Cordilleran Cambrian strata become better known, numerous new formations doubtless will be designated, especially in the regions distant from the geosynclinal axis. Nevertheless, in view of the widespread continuity of the formations involved and in the interest of unification of stratigraphic nomenclature, we have prepared the following descriptions and designations of the Cambrian rocks of western Grand Canyon. (see Fig. 3).

PROSPECT MOUNTAIN QUARTZITE

The basal Cambrian strata of western Grand Canyon and vicinity are assigned to the Prospect Mountain quartzite. These rocks range in color from dark red-brown through brown and buff to cream. They are mostly coarsely granular, somewhat feldspathic (especially near the base), locally glauconitic, and mostly cross-laminated. The basal 118 feet are massively cross-bedded ($\frac{1}{2}$–3 feet thick), but the upper 61 feet are more thinly bedded (3–8 inches), with green or red

[5] E. T. Schenk and H. E. Wheeler, "Cambrian Sequence in Western Grand Canyon, Arizona" (abstr.), *Bull. Geol. Soc. Amer.*, Vol. LII (1941), pp. 1983–84.

shale intercalations and partings. Gilbert[6] and Wheeler and Kerr[7] include these upper, thinly bedded sandstones with the overlying shale. However, since these strata are predominantly quartzitic sandstones, we follow C. R. Longwell,[8] who, in the near-by Virgin Range, incorporates them with the underlying quartzites, with which they are more closely allied lithologically.

The type locality of the Prospect Mountain formation is in the Eureka district, Nevada, 250 miles northward, where it was defined by Arnold Hague.[9] This formation has been identified elsewhere in Nevada at Pioche[10] and Delamar[11] and in Utah in the House Range[12] and Gold Hill district.[13] Wheeler[14] has traced the Prospect Mountain quartzite southward from Eureka to western Grand Canyon. Between these two localities it has been observed at average intervals of about 42 miles (Cave Valley, Pioche, Delamar, Mormon Range, and Virgin Mountains).

Wheeler and Kerr[15] assign these basal western Grand Canyon sediments to the "Tapeats"[16] sandstone of the Granite Gorge area of the Canyon. However, since the Prospect Mountain and "Tapeats" formations are lithogenetically the same, the former name is employed because of its priority of definition.

[6] *Op. cit.*, p. 163.

[7] *Op. cit.*, p. 7.

[8] "Geology of the Muddy Mountains, Nevada, with a Section through the Virgin Range to the Grand Wash Cliffs, Arizona," *U.S. Geol. Surv. Bull. 798* (1928), p. 23.

[9] "Abstract of Report on the Geology of the Eureka District, Nevada," *U.S. Geol. Surv., 3d Ann. Rept.* (1883), pp. 253–55.

[10] L. G. Westgate, "Geology of Pioche, Nevada and Vicinity," Part I: "General Geology," *Trans. Amer. Inst. Min. Met. Eng.*, Vol. LXXV (1927), pp. 816–28; and "Geology and Ore Deposits of the Pioche District, Nevada," Part I: "General Geology," *U.S. Geol. Surv. Prof. Paper 171* (1932), pp. 1–44.

[11] Eugene Callaghan, "Geology of the Delamar District, Lincoln County, Nevada," *Univ. Nevada Bull.*, Vol. XXXI, No. 5 (1937), pp. 15–17.

[12] C. D. Walcott, "Nomenclature of Some Cambrian Cordilleran Formations," *Smith. Misc. Coll.*, Vol. LIII, No. 1804 (1908), pp. 1–12.

[13] T. B. Nolan, "The Gold Hill Mining District, Utah," *U.S. Geol. Surv. Prof. Paper 177* (1935), pp. 4–6.

[14] *Loc. cit.* [15] *Op. cit.*, pp. 6–7.

[16] L. F. Noble, "The Shinumo Quadrangle, Grand Canyon District, Arizona," *U.S. Geol. Surv. Bull. 549* (1914), pp. 61–62.

The age of detrital strata as widespread as those of the Prospect Mountain formation may be expected to vary appreciably from place to place. Wheeler[17] has stated that the age of these rocks ranges from latest pre-Cambrian(?) and earliest Cambrian in eastern Nevada to early medial Cambrian in the Granite Gorge area of the Grand Canyon. In the western Grand Canyon this 179-foot sequence of rapidly deposited quartzites lies with perfect conformity immediately below the Lower Cambrian *Olenellus* zone of the Pioche shale. The entire formation in this area, therefore, is here assigned to the Lower Cambrian series.

The nearly horizontal Prospect Mountain quartzite lies unconformably upon granite gneiss assigned to the Archean by Ian Campbell and J. H. Maxon.[18] These basement rocks and most of the Prospect Mountain quartzite in this area are now submerged beneath the waters of Lake Mead.

PIOCHE SHALE

The type locality of the Bright Angel shale[19] is near the mouth of Bright Angel Canyon in the Granite Gorge area of the Grand Canyon, where the formation conformably overlies the "Tapeats" sandstone. Wheeler and Kerr[20] described the Bright Angel, together with the "Tapeats" (Prospect Mountain), in western Grand Canyon. We agree that the lithogenetic counterpart of the Bright Angel formation is present in this region; but, as stated in a previous paragraph, its equivalent in the Pioche, Nevada, district has been logically and successfully subdivided into the Pioche shale, Lyndon limestone, and Chisholm shale.[21] These three formations are lithologically, stratigraphically, and faunally traceable to western Grand Canyon,[22] where their recognition argues against the continued local use of the term "Bright Angel." Continued employment of the term in the Grand Canyon area beyond the eastern limit of the Lyndon formation is not questioned.

[17] *Loc. cit.*

[18] "Geological Studies of the Archean Rocks at Grand Canyon, Arizona" (abstr.), *Carnegie Inst. Wash. Year Book 37* (1938), pp. 359–64.

[19] Noble, *op. cit.*, pp. 62–64.

[20] *Op. cit.*, pp. 8–9.

[21] Westgate, "Geology and Ore Deposits . . . . ," *op. cit.*

[22] Wheeler, *loc. cit.*

In western Grand Canyon the Pioche shale lies conformably upon the Prospect Mountain quartzite. At the measured section it consists of 304 feet of chocolate to red-brown, buff, cream, and light-green micaceous shales and sandstones. Some of the sandstones are cross-bedded, and some are locally glauconitic.

The Pioche shale of Walcott,[23] at its type locality 130 miles northward, has been shown by Charles Deiss[24] to consist of a lower 600-foot portion of early Cambrian age and an upper part (370 feet) belonging to the Middle Cambrian. Deiss restricted the Pioche formation to the Lower Cambrian strata and applied the term "Comet shale" to the upper 370 feet. H. E. Wheeler and D. M. Lemmon[25] rejected the revision because the "Pioche" and "Comet" shales, so delimited, are not valid cartographic units, even within the Pioche district. These shales are now known to occur from Eureka to the Grand Canyon,[26] cropping out in an area of many thousands of square miles. Throughout this area the Lower-Middle Cambrian boundary follows no recognizable line of lithologic or formational demarcation. In fact, this time boundary is known to transect at least one formational contact.

Thus, the Pioche shale, originally defined by Walcott,[27] is present in western Grand Canyon. As in the type area, the Pioche formation contains both Lower and Middle Cambrian strata. The exact position of the boundary, however, is unknown. The Lower Cambrian *Olenellus* zone occurs about 13 feet above the base (*F-15*), whereas a species of the lower Middle Cambrian genus *Albertella* occurs 31 feet below the top (*F-44*). This leaves 260 feet of strata within which the boundary may be allocated approximately by interpolation. E. D. McKee[28] states that the undetermined interval is reduced to 55 feet

[23] *Loc. cit.*

[24] "Cambrian Formations and Sections in Part of Cordilleran Trough," *Bull. Geol. Soc. Amer.*, Vol. XLIX (1938), pp. 1158–60.

[25] "Cambrian Formations of the Eureka and Pioche Districts, Nevada," *Univ. Nevada Bull.*, Vol. XXXIII, No. 3; "Geol. and Min. Ser. No. 31," (1939), pp. 34–35.

[26] Wheeler, *loc. cit.*

[27] *Loc. cit.*

[28] "Studies on the History of Grand Canyon Paleozoic Formations" (abstr.), *Carnegie Inst. Wash. Year Book 38* (1939), pp. 313–14.

of sediments. Presumably this announcement was made on the basis of an erroneous field identification of olenellid spines by the writers, and an occurrence of the Middle Cambrian genus *Anoria* collected by McKee and identified by C. E. Resser, about 73 feet below the top of the shale (*F. C. 761*). If, however, the *Anoria* specimen has been correctly identified, here is an instance of a species of the genus *Anoria* occurring 42 feet stratigraphically below an occurrence of *Albertella*. Such a case, perhaps, is not impossible, though it is a reversal of their normal relative positions. If this be true, the interval of undetermined age is now 218 feet. The approximate series boundary is established arbitrarily and provisionally at the top of the dark-maroon sandstone 186 feet above the base of the Pioche shale.

LYNDON FORMATION

The Pioche shale in western Grand Canyon, as at Pioche, Nevada, is overlain conformably by the Lyndon formation. At its type locality in Lyndon Gulch on the west side of the Highland Range, Pioche district, the Lyndon[29] formation consists of 345 feet of limestones. By the lateral gradation of the upper beds into shale, the formation thins gradually in a southerly direction until it is reduced to a thickness of only 45 feet at Whitney Ranch in the Virgin Mountains, about 25 miles north of the Canyon. This is the cliff-forming limestone described by Longwell[30] as a part of the "Bright Angel shale(?)." Southward from Whitney Ranch a tongue of detrital sediment separates the limestone into two members more or less equal in thickness. In addition, these calcareous members are secondary (epigenetic) dolomites in the Lake Mead region. In western Grand Canyon the Lyndon formation is represented by a lower 12-foot member of rust-colored, massive dolomite; a middle, 40-foot member of shale, sandstone, and dolomite; and an upper 18-foot dolomite member which is identical in appearance to the basal layer. These three distinct members are the units *b*, *c*, and *d* of the "Tonto shale," briefly described by Gilbert.[31]

The Lyndon formation is characteristically a cliff-forming unit, in

[29] Westgate, "Geology and Ore Deposits . . . . ," *op. cit.*

[30] *Loc. cit.*

[31] *Op. cit.*, p. 163.

contrast with the slope-forming shales above and below. Neither here nor elsewhere is the Lyndon formation known to contain diagnostic fossils. Nevertheless, since it lies conformably between the lower Middle Cambrian Pioche and Chisholm shales, the Lyndon also belongs to the same series. The occurrence of a species of the genus *Albertella* just 31 feet below its base and of both *Glossopleura* and *Anoria* in the overlying Chisholm shale indicates the *Zacanthoides-Anoria* zone as the approximate "time-stratigraphic" position of the Lyndon formation in western Grand Canyon.

CHISHOLM SHALE

As stated previously, the Chisholm shale is the uppermost of the three formations which together constitute the lithogenetic equivalent of the Bright Angel shale in western Grand Canyon and vicinity. Like the Pioche and Lyndon formations, the type section of the Chisholm is also in the Pioche district. In the general vicinity of the type locality the Chisholm shale has an average thickness of about 100 feet. Although its thickness does not remain constant toward the south (Callaghan[32] reports 300 feet at Delamar, Nevada), we have measured a 99-foot sequence in western Grand Canyon. In this region the Chisholm strata are mostly maroon-red, dark-brown, and dark-green shales and sandstones with a minor amount of limestone and dolomite.

The Chisholm trilobite fauna in this area include the following: *Alokistocare althea* Walcott, *Anoria* n. sp., *Clavispidella enucleata* Resser, *Elrathia* sp., *Glossopleura mckeei* Resser, and *Glyphaspis* sp. All these genera except *Anoria* and *Elrathia* have been reported from the Chisholm shale in its type area. In fact, *Zacanthoides* is the only characteristic, type Chisholm trilobite genus which is absent from this faunal list.

Two of these western Grand Canyon Chisholm fossils are conspecific with forms originally defined from the Granite Gorge region. *Alokistocare althea* and *G. mckeei* occur in the lower Bright Angel shale at Indian Garden Springs (opposite Bright Angel Canyon) and in the uppermost "Tapeats" (Prospect Mountain) formation in

[32] *Op. cit.*, p. 17.

Nankoweap Valley. The genus *Anoria*, represented by *A. tontoensis* (Walcott), is also found at these localities.

The duration of the period of eastward advance of the sea from the Grand Wash Cliffs region to the Granite Gorge area is well illustrated by these biostratigraphic relationships. The Prospect Mountain–Pioche contact, in western Grand Canyon, as stated previously, belongs approximately to the lower Cambrian *Olenellus* zone; while its equivalent lithogenetic "horizon" in the Granite Gorge region (the "Tapeats"–Bright Angel contact) belongs to the *Anoria* or *Glossopleura* zones of the Middle Cambrian. Further, the first strata of Bright Angel deposition in the Granite Gorge region are contemporaneous with the last stages of sedimentation of its lithogenetic counterpart in the Lake Mead area (Chisholm shale).

Together with this faunal and formational data, to indicate the gradual eastward migration of Cambrian facies in the Grand Canyon region, McKee[33] is now preparing much additional evidence, including many detailed sections in the region, to show the lateral gradation of the various strata and members of the sequence.

PEASLEY LIMESTONE

The same supra-Chisholm limestone is continuous from the Grand Canyon to Pioche, Nevada, and beyond. H. E. Wheeler[34] applied the name "Peasley limestone" to these rocks in the Pioche district. Schenk and Wheeler,[35] believing the Peasley formation to be the exact lithogenetic equivalent of the Muav limestone of the Granite Gorge area, accepted the latter on the basis of its priority of definition. However, McKee[36] has shown that the upper strata of the type Muav limestone carry into western Grand Canyon, where they disconformably overlie limestones (Mead formation) which in turn rest disconformably upon strata lithogenetically equal to the lower type Muav. In other words, if the type Muav were carried westward down the Canyon to the Lake Mead region, portions of it would

[33] Verbal communication, 1941.

[34] "Revisions in the Cambrian Stratigraphy of the Pioche District, Nevada," *Univ. Nevada Bull.*, Vol. XXXIV, No. 8; "Geol. and Min. Ser. No. 34" (1940), pp. 17–27.

[35] *Loc. cit.*

[36] Verbal communication, 1941.

occur above and other portions below the Mead formation. Except for the disconformities involved, this nomenclatural problem is similar to that of the Bright Angel shale and its equivalents in the western part of the Canyon.

In view of the disconformities within its western formational equivalents, the name "Muav" cannot be employed in the section herein described. The supra-Chisholm limestones of this region, as stated above, are called "Peasley" at Pioche; and, since these rocks are continuous between Pioche and western Grand Canyon, the name "Peasley" is applicable in both regions.

In the Pioche district the Peasley limestone consists mostly of dark-gray, medium-grained, thickly bedded to massive limestone. It weathers with dull-gray surfaces, showing irregular banding or mottling in zones of intercalated impurities or partial secondary dolomitization. Measured thicknesses in the Pioche district vary between 120 and 175 feet in consequence of the post-Peasley erosion interval. Less of the formation is removed toward the south. It measures about 400 feet in the Lake Mead region. Here the upper half of the Peasley contains three members with thinly bedded limestones, shales, and arenaceous beds. These detrital layers are less resistant to erosion and form the characteristic slopes, in contrast with the cliff-forming limestones with which they are interbedded.

The following trilobite genera occur near the base of the Peasley formation in western Grand Canyon: *Alokistocare*, *Chancelloria*, *Glossopleura*, and *Kootenia*. *Elrathia* is found near the middle of the formation. The upper beds carry *Elrathia*, *Kootenia*, and *Solenopleurella*. These are Middle Cambrian genera that appear to belong in part, at least, to the *Glossopleura-Kootenia* zone.

MEAD LIMESTONE

A variable thickness of white to pale-pink, fine-grained to dense, somewhat mottled, massive limestone lies disconformably upon the Peasley limestone in this region. On the basis of lithologic character, stratigraphic position, and disconformable relationships, this formation fits the description of the undolomitized phase of the Burrows formation of the Pioche district.[37]

[37] Wheeler, "Revisions in the Cambrian . . . . ," *op. cit.*, pp. 27–28.

However, because the Burrows has not yet been recognized in the southern half of the intervening distance between Pioche and the Grand Canyon, we are reluctant to apply the name in the Lake Mead region. Therefore, the name "Mead limestone" is proposed for these strata.

The disconformity (Fig. 4) is readily recognized where accessible, the contact usually being manifested by a marked recession in the

FIG. 4.—Disconformity between Peasley limestone (*below*) and Mead limestone (*above*) on south wall of Grand Canyon about 1¼ miles east of Grand Wash Cliffs, Arizona.

cliff. Detrital calcareous sediment is present from several inches to a few feet in thickness. A maximum of 15 feet of relief was observed in the topmost beds of the Peasley limestone.

These limestones have a thickness of 110 feet at the measured section, although the thickness is quite variable in the surrounding region. This variation is due to a post-Mead disconformity, which has completely removed the formation a short distance to the eastward.

No diagnostic fossils have been found in the Mead formation.

Therefore, its age is not determinable at present. Nevertheless, that it belongs to the Cambrian is probable because of its position between the underlying Middle Cambrian Peasley limestone and the overlying strata, which are similar in lithology and stratigraphic position to those carrying Upper Cambrian faunas in near-by regions. If the Mead limestone is the equivalent of the Burrows formation, as appears likely, it belongs quite certainly to the Middle Cambrian series.

CAMBRIAN(?) AND POST-CAMBRIAN DEPOSITS

Disconformably above the Mead formation lies a thick series of dolomites and limestones, which are in turn covered by limestones of Mississippian age. The gray dolomites and white laminated limestones occurring immediately above the Mead formation are at least higher in the Middle Cambrian series than any unit known elsewhere in the Canyon and may actually be of Upper Cambrian age or younger. From about 1,000 feet above the base of this undifferentiated calcareous sequence L. L. Sloss[38] has identified corals as belonging to the genus *Streptelasma*. Although these specimens are poorly preserved, they nevertheless resemble Devonian species of that genus.

The top of the Cambrian sequence in western Grand Canyon, therefore, cannot be definitely located until diagnostic fossils are found in this thick series of dolomites and limestones.

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[38] Personal communication, 1939.

TABLE 1

CAMBRIAN SECTION OF WESTERN GRAND CANYON*

		Feet
Devonian(?) and Cambrian	Dolomite and limestone: Mostly light-gray to almost white, medium-grained, laminated to thickly bedded dolomite and limestone, weathering buff to reddish-brown. Forms cliff composed of series of receding ledges. *Streptelasma* sp. occurs about 1,000 feet or more above the base of this sequence†	
	Disconformity	
Middle(?) Cambrian	Mead limestone (type locality): Light-gray to pink, fine-grained, laminated to thin-bedded (outcrops massive), somewhat mottled limestone. Forms cliff. The post-Mead unconformity exhibits about 50 feet of local relief at type locality. At some places, however, entire formation has been removed	0–110
	Disconformity	
Middle Cambrian	Peasley limestone	
	Limestone and shale: Medium to dark-gray and tan, thin-bedded (platy) limestone, intercalated with thin partings of green, paper-thin shale and buff to greenish-gray, micaceous, arenaceous shale. Locality *F-40* (*Kootenia* n. sp. and *Solenopleurella* n. sp.), 55 feet above base. Locality *F-47* (*Elrathia* n. sp.), about 40 feet above base. Forms 35° slope	75–90
	Edgewise conglomerate: Thin slabs of buff or brown, argillaceous limestone standing at various angles	1
	Talus-covered slope: 35° slope, covered with slabs of gray to buff, thinly bedded, platy limestone talus	8
	Limestone: Gray and tan mottled, fine- to medium-grained limestone with numerous irregular silty laminae etched out on weathered surfaces. Forms massive, sheer cliff	42
	Limestone and shale: Gray, fine-grained to dense, thin-bedded (often irregularly bedded), platy limestone, intercalated with greenish shale and siltstone layers ½–2 inches thick. Forms 25° slope	18
	Limestone: Gray, fine-grained to dense, thinly and irregularly bedded limestone with irregular, brown, arenaceous partings etched out on weathered surfaces. Forms cliff	66
	Limestone and shale: Gray, fine-grained to dense, thinly bedded, platy, limestone, interbedded with green to buff, micaceous and arenaceous shale. Locality *F-37* ((*Hyolithes* sp. and *Elrathia* n. sp.) in uppermost beds. Forms 23° slope	16

* From the south canyon wall about 1¼ miles east of Grand Wash Cliffs.

† The *Streptelasma* locality is on Grand Wash Cliffs, about a mile north of the western terminus of the Grand Canyon.

TABLE 1—*Continued*

		Feet
Middle Cambrian (*continued*)	Peasley limestone (*continued*)	
	Limestone: Medium- to dark-gray, fine-grained to dense limestone with numerous, dark-brown irregular, silty laminae etched out on weathered surfaces. Forms sheer cliff	133
	Limestone: Medium-gray, thin-bedded (½–3 inches) limestone intercalated with buff to brown, argillaceous limestone (gives mottled appearance). Locality *F-39* (*Hyolithes* sp., *Alokistocare* sp., *Chancelloria* cf. *eros* Walcott, *Glossopleura* sp., and *Kootenia* sp.). Usually forms slight recess at base of cliff	29
Total		403
	Chisholm shale	
	Shale and sandstone: Maroon and dark-brown, mostly paper-thin shales, alternating with thin beds of buff to dark-brown sandstone. A few thin beds of gray limestone occur in upper part; and a few thin, rust-colored dolomite strata in lower part. The following fossil localities have been found in the Chisholm shale:	
	F-16, 68 feet above base; *Alokistocare althea* (Walcott), *A. lepidum* Resser, *Clavispidella enucleata* Resser, *Elrathia* n. sp., and *Glyphaspis* sp.	
	F-17, 44 feet above base; *Lingulella* n. sp., *Margaritia?* sp., *Alokistocare* cf. *althea* Walcott, *Anoria* n. sp., *Elrathia* sp., and *Glossopleura mckeei* Resser	
	F-38, 30 feet above base; *Alokistocare?* Forms 30° slope	99
	Lyndon formation	
	Dolomite: Light-brown, massive dolomite, weathering to rust color. Forms sheer cliff	18
	Shale: Light-green, arenaceous shale, with few intercalations of dark-gray and rust-colored dolomite. Forms blocky cliff or slight recess	29
	Sandstone: Pale-green, buff, and dark-brown sandstones, containing worm borings. Forms slight recess	11
	Dolomite: Light-brown, massive dolomite, weathering to rust color. Forms sheer cliff	12
Total		70
	Pioche shale	
	Shale and sandstone: Light-green to buff, paper-thin shales interbedded with green and buff, fine-grained, quartzitic sandstones. Trilobite tracks and worm borings abundant. Locality *F-44* (*Albertella* n. sp.), 67 feet above base. Locality *F.C. 761* (*Anoria* sp.) about 25 feet above base. Forms 30° slope	98

TABLE 1—*Continued*

		Feet
Middle Cambrian (*continued*)	Pioche shale (*continued*)	
	Sandstone: Light-gray, medium-grained, quartzitic, somewhat cross-laminated sandstone. Forms low ledge	2
	Sandstone: Red to buff, fine-grained, partly micaceous sandstones. Forms 25° slope	18
Lower Cambrian (boundary approximate)	Sandstone and shale: Dark-maroon-red, fine- to medium-grained, thin- to thick-bedded (some cross-bedded) sandstone with few thin beds of paper-thin shale. Uppermost 10 feet forms red cliff; remainder forms 30° slope	34
	Shale and sandstone: Green to buff, micaceous, papery shale, intercalated with thin beds of quartzite and sandstone. Forms 30° slope	15
	Shale and sandstone: Red, gray, and pale-green, micaceous shales, alternating with thin beds of light-cream, quartzitic sandstones. Shales carry abundant trilobite tracks and worm borings. Forms 24° slope	55
	Sandstone: Light-gray, coarse-grained, sandstone (lower part cross-laminated). Forms ledge	8
	Sandstone and shale: Buff to dark-brown and gray, coarse-grained, quartzitic sandstones, interbedded with pale-green to deep-red, paper-thin shales. Shale more abundant in uppermost 30 feet. Numerous worm borings throughout. Locality *F-15* (*Antagmus* sp., *Nisusia* sp., and *Olenellus* sp.) 13 feet above base. Forms 20° slope	74
Total		304
	Prospect Mountain Quartzite	
	Quartzite and shale: Light-cream, coarse-grained, thinly and distinctly bedded (3–8 inches) quartzite with partings of pale-green to buff shale. Forms steep, steplike slope	15
	Quartzite and shale: Buff, brown to dark red-brown, medium- to coarse-grained, thinly bedded (3–12 inches) quartzite and quartzitic sandstones with shale partings. Forms steep steplike slopes	46
	Quartzite: Cream, buff, brown to dark red-brown, coarse-grained, thickly bedded to massive, partly cross-bedded quartzite. Forms massive sheer cliff	118
Total		179
	Unconformity	
	Archean	
	Granite gneiss injected by numerous pegmatite dikes	

THE GENESIS OF SUBORDINATE CONJUGATE FAULTING IN THE KERN RIVER SALIENT

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ABSTRACT

All oil pools on the Kern River salient, which produce from Temblor (Miocene) formations, are fault trapped. These "trap" faults are normal, with the downthrow side to the east, and are found opposite prominent fault scarps in the Sierra Nevada massif. The faults of the massif are also normal, of considerably larger displacement than the trap faults, and dip westward so that a graben-like structure is formed. Because the extrapolated fault planes are found to intersect downward at a depth of about 3 miles and because the trap faults and those of the massif die out similarly along the strike, it is suggested that the trap faults are dependent upon the latter for their development. It is further suggested that such *subordinate conjugate* faults form by movement along sets of shears resulting from a rotational force couple.

INTRODUCTION

A few miles southeast of Bakersfield, California, the granodiorite of the Sierra Nevada massif extends bluntly wedgelike into the sediments of the San Joaquin Valley. The Kern River has cut a deep canyon in the southern, and most prominent, part of the wedge, which has resulted in its being known locally as the "Kern River salient." In the sediments the salient is reflected as a gently plunging nose, and all the "east side" oil pools north of the Kern River are structurally related to it. These include Round Mountain, Coffee Canyon, Pyramid Hill, Tampico, Mount Poso, Mount Poso (Ring), North Mount Poso, Kern Front, Premier, Fruitvale, and Kern River pools. The oil, with the probable exception of that of the Kern River pool, is trapped by faults.

Because of the important production numerous stratigraphic and paleontologic studies have been carried on in this area since the discovery of the Kern River pool in 1900. Structural studies, however, have been confined essentially to productive areas, with a consequent restriction of published data.

The genesis of the faulting associated with the salient has been considered in two reports. L. S. Fox[1] explained it as block faulting

[1] "Structural Features of the East Side of the San Joaquin Valley," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XIII (1929), pp. 101–8.

due to tension. R. D. Reed and J. S. Hollister[2] in 1936 tentatively assumed the faulting to have resulted from opposing plunges of the Sierra Nevada and the San Joaquin Valley, considering these as simple folds, with the faults located in the hinge area. Alex Diepenbrock's reports[3] on the Mount Poso and Round Mountain fields, while excellent discussions of the stratigraphy and structural relationships, do not consider the origin of the faulting.

FIG. 1.—Index map showing location of area and major regional features

GENERAL GEOLOGY

As shown on the index map of Figure 1, the Kern River salient lies within two geologic provinces: the Sierra Nevada and the San Joaquin Valley. Here, Cenozoic marine and nonmarine detritals overlap the erosional surface of the westward-tilted block of Mesozoic granodiorite. From wells drilled through this sedimentary blanket

[2] "Structural Evolution of Southern California," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XX (1926), p. 1617.

[3] "Mt. Poso Oil Field," *California Oil Fields*, Vol. XIX, No. 2 (1933), pp. 5–36; "Round Mountain Field," *ibid.*, No. 4 (1934), pp. 5–20.

it has been determined that the underlying crystalline rock is a continuation of the Sierran massif, thus precluding the possibility that the Franciscan, a much more mobile-type "basement," occurs as far east as this area. The crystalline rock is pre-Tertiary, probably late Jurassic, and the sedimentary contact strikes approximately N. 15° W., roughly paralleling the trend of the main mountain mass.

North of the Tulare County line, Pliocene and Pleistocene deposits overlap the older formations and lie directly on the crystallines. Southward, however, Miocene formations appear, and east of Mount Poso they form a belt 3 miles wide, which increases to 6 miles at the Kern River. West of this belt the Pliocene and Pleistocene deposits are separated from the Miocene by the Kern River unconformity and crop out basinward until covered by recent alluvium. Both series dip gently westward, the Miocene somewhat more steeply than the Pliocene and Pleistocene.

The most prominent topographic features of the area are the scarp of the Kern River fault and the canyons of Kern River and Poso Creek. These canyons are cut in the crystalline rock of the Sierran mass and are indicative of relatively recent uplift. Subdued topography is characteristic of the granitoid upland, and remnants of the old Sierran erosional surface exist but slightly modified at numerous sites. Thus, since the last major crustal movements, erosion is still in the youthful stage in this resistant rock.

Westward, in the less resistant sediments, degradation has reduced the topography to the stage of regional maturity. No pronounced interfluve areas remain; and below the points where Kern River and Poso Creek debouch from their precipitously walled canyons, flood plains have been built along their slightly meandering courses. Faulting has played a significant role in the location of some streams and washes, and portions of the courses of both Poso Creek and Granite Canyon are subsequent. Near the contact, streams in the overlap zone have been incised more deeply. This probably resulted from local movements in the basement.

REGIONAL STRUCTURE

The Sierra Nevada constitutes a single mountain mass somewhat over 400 miles long and averaging about 70 miles in width. The

trend of the northern three-fourths is approximately N. 30° W., while from Mount Whitney south to Tehachapi Creek the form is arcuate, convex eastward, with the chordal trend roughly north-south. From Tehachapi Creek to the junction with the Coast Ranges the Tehachapi Mountains swing southwestward, with an overall trend of approximately N. 45° E.

The Sierra as a whole is a large westward-tilted fault block, composed mostly of granitic-type rocks, with the eastern margin delineated by the Sierra fault scarp and the western margin covered by the sedimentary mantle of the San Joaquin Valley. From Mount Whitney, the highest part of the block, the tilt is southward as well as westward, and the defining fault on the east is not of so great structural and topographic significance.

The Tehachapi Mountains, although an extension of the main Sierran massif, should probably be considered a structural subprovince. They are bounded on the south and east by the Garlock fault, have risen as a block, and dip northwestward to the San Joaquin Valley. Dips increase as the Coast Ranges are approached; near the transitional junction with the San Emigdio Mountains the major folds in the overlapping sediments are all overturned to the north.

It is somewhat conjectural as to when the boundary faults were initiated. Waldemar Lindgren[4] suggested that the Sierra fault was formed during late Cretaceous and that intermittent movement occurred throughout Cenozoic time. More recent work, however, by G. D. Louderback,[5] Adolph Knopf,[6] F. E. Matthes,[7] and others has led to the conclusion that the uplifts during late Cretaceous and throughout the Tertiary were a result of regional doming and that the boundary faults were not formed until late Pliocene or early Pleistocene, the period of maximum uplift. This conclusion is based

[4] "Tertiary Gravels of the Sierra Nevada," *U.S. Geol. Surv. Prof. Paper 73* (1911), pp. 41–43.

[5] "Period of Scarp Production in the Great Basin," *Univ. Calif., Dept. Geol. Sci. Bull.*, Vol. XV (1924), pp. 1–44.

[6] "Geological Reconnaissance of the Inyo Range and the Eastern Slope of the Southern Sierra Nevada," *U.S. Geol. Surv. Prof. Paper 110* (1918), pp. 86–92.

[7] "Geological History of the Yosemite Valley," *U.S. Geol. Surv. Prof. Paper 160* (1930), pp. 28–30.

on a correlation of the Sierran flows with marine and nonmarine sediments on the east and west sides of the Sierra; on information derived from a study of ancient stream profiles; and geomorphic studies involving the development of the scarps, the different erosion levels, and the migration of the divide eastward. It should be noted that R. W. Webb[8] thought that the Kern Canyon fault, which strikes north-south north of Kernville and roughly bisects the southern Sierra, was pre-Pliocene.

Along the western margin of the block a number of scarps, which extend from the Yosemite region southward to the Kern River, have been described by B. F. Hake.[9] With a few minor exceptions these faults are normal, with the downthrow side toward the valley. They are probably a result of a force couple—the uplift of the Sierra to the east and the downwarp of the valley to the west. Thus, the western side of the massif is partially bounded by faults, along which much less movement has occurred than along the Sierra fault.

In the area of the Kern River salient two such scarps are found, both striking about N. 30° W. Southward, about halfway between the Kern River and the Tehachapi Mountains, the scarps bordering Bear Mountain and the Tejon Canyon fault also parallel this trend. In the Tehachapis proper, H. W. Hoots[10] found that the "deep canyons show a striking tendency toward parallelism, all of them trending in a northwest direction parallel to the two known cross faults of the Tehachapi foothills." He concludes that this alignment is a result of fracturing and that "these fractures are potential shear zones developed in crystalline rocks north of the San Andreas fault as a result of thrusts originating farther south." Thus it would appear that the effects of the diastrophism of the Tehachapis may have extended as far north as the Kern River region and that the structural development of the Kern River salient was influenced by the tectonics of the Sierra and Tehachapis. This is borne out by an increasing structural complexity as the Tehachapis are approached.

[8] "The Kern Canyon Fault," *Jour. Geol.*, Vol. XLIV (1936), pp. 631–38.

[9] "Scarps of the Southwestern Sierra Nevada," *Bull. Geol. Soc. Amer.*, Vol. XXXIX (1928), pp. 1017–30.

[10] "Geology and Oil Resources along the Southern Border of the San Joaquin Valley," *U.S. Geol. Surv. Bull. 812-D* (1930), pp. 301–19.

STRUCTURE OF THE KERN RIVER SALIENT

The block diagram of Figure 2 shows the structure of the trap-fault area. Contours are on top of the Vedder (Miocene) formation and are based on subsurface data. Between the sedimentary-igneous contact and the contoured Vedder the block is shown as stripped to the basement. This was necessary because of paucity of subsurface data.

The Kern River oil field is located on the nose which is the chief expression of the salient valleyward. The nose plunges to the southwest, and north of it the strike of the formations is about N. 20° W.; the Miocene beds dip westward at 6°, and the younger formations at 4°. South of the nose the strike changes to nearly northwest, and the dips increase a few degrees. This change in trend and increase of dip indicate a depositional re-entrant to the south, where there are 3,700 feet of Temblor, in contrast to only 2,800 feet at Round Mountain.

A majority of the faults of the salient are longitudinal, paralleling the regional grain, and are more numerous in the southern part of the area, opposite the scarps of the Kern River and Poso Creek faults. There are also a few small cross faults which strike about N. 70° W. The fault pattern corresponds well with the joint pattern of the massif, which in turn matches the trends found in the Mount Lyell–Mount Whitney section of the Sierra by Evans B. Mayo.[11] He concluded that the orientation of structures in that region was a result of stresses active throughout the batholithic emplacement. It is then possible that the faults in the salient area lie in the pre-existing zones of weakness, the "joint pattern," and that the effect of stresses from the Tehachapis may be secondary this far north.

BOUNDARY FAULTS

As stated by Fox,[12] all faults in the area are normal. Of these, the Kern River fault, which has been described by G. K. Gilbert,[13] Eliot

[11] "Deformation in the Interval Mt. Lyell–Mt. Whitney," *Bull. Geol. Soc. Amer.*, Vol. LII (1941), pp. 1001–84.

[12] *Op. cit.*

[13] "Studies in the Basin Range Structure," *U.S. Geol. Surv. Prof. Paper 153* (1928), pp. 86–89.

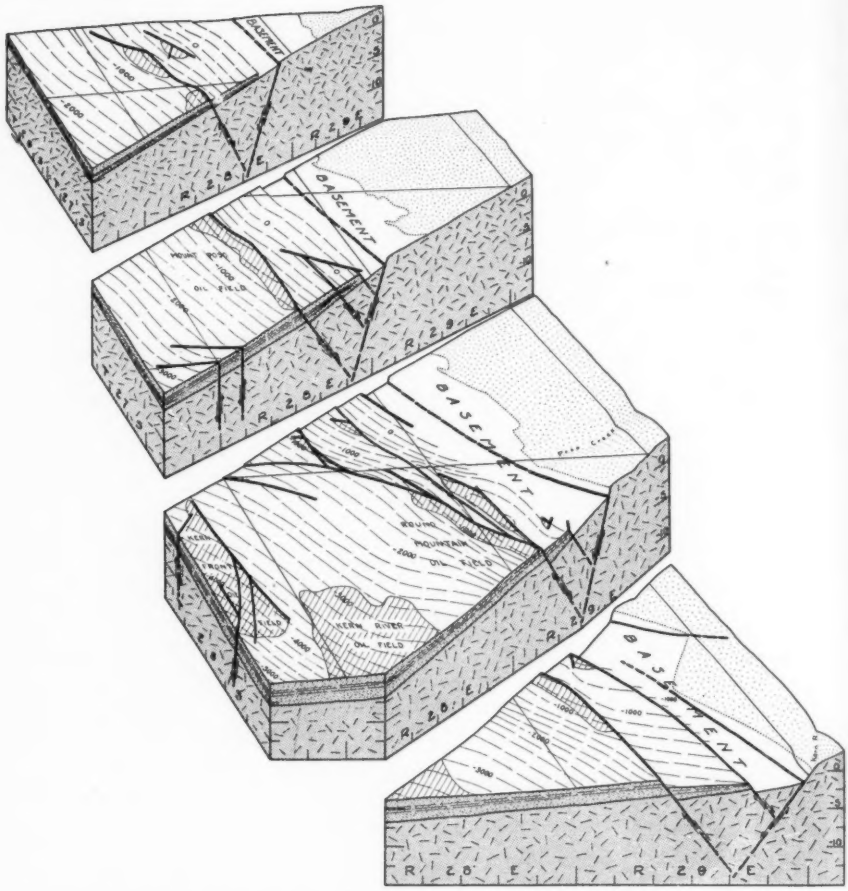

FIG. 2.—Block diagram showing structural relationships. Contours on top of Vedder. Area between contoured Vedder and sedimentary-igneous contact shown as stripped to basement. Productive area crosshatched.

Blackwelder,[14] and others, is the most prominent topographically. The Kern River, in cutting its gorge, has sapped the easily eroded clastics and exposed, as well as modified, the fault surface. The scarp, where the Kern leaves the crystallines, is over 1,000 feet high, the lower 400 feet of which apparently represents the old fault surface. The upthrow side is formed by the granodiorite of the Sierra, and the downthrow side by Tertiary sediments. This fault strikes N. 30° W. and dips southwestwardly at about 60°. Northwestward it may be traced from surface indications for some 3 or 4 miles, while southeastward it soon becomes obscured in the complexities of the massif. A well drilled a short distance from the scarp reached a depth of over 1,000 feet without penetrating the basement. Thus, it is certain that the throw is at least 2,000 feet.

The Poso Creek fault is about 4 miles north of the Kern River fault. Its scarp can be traced northwestward several miles to Poso Creek, beyond which it disappears under the sediments of the valley. Northward extension of this fault is shown as dashed because of uncertainty as to its exact position. From airplane photographs B. L. Clark[15] has suggested that the Kern River fault may extend northwestward, under the sediments, for 20 miles. If this is correct, the Poso Creek fault must either die out rapidly, join, or intersect the proposed Kern River fault extension. The rapid dying-out, however, of the fault which parallels the Kern River fault immediately to the west and the structural contiguity of the re-entrant formed by the Kern River and Poso Creek faults (Fig. 2) and of the Pyramid Hill area suggest that the Kern River fault also rapidly disappears northward, neither joining nor intersecting the Poso Creek fault.

Upon extending the sediment-covered basement profiles of the block diagram (Fig. 2) it is evident that they fail to meet the surface profiles of the granodiorite. This discrepancy is over 1,000 feet on the southernmost profile crossing the Poso Creek fault and, as expected, decreases northward. Since the sediments were deposited on a well-developed erosion surface, this anomaly is best explained by

[14] "Scarp at the Mouth of the Kern River Canyon," *Bull. Geol. Soc. Amer.*, Vol. XXXVIII (1927), p. 207.

[15] "Tectonics of the Valle Grande," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XIII (1929), pp. 199–238.

faulting. Further, since the crystallines east of the overlap have been weathered and eroded, the actual displacement must have been greater than that shown. In the northeast and east-central rectangles of the Bakersfield quadrangle more deeply incised canyons and an abrupt change in elevations on profiles constructed in an east-west direction afford additional evidence of basement faulting of considerable magnitude. Thus the northward extension of the faulting seems adequately substantiated, although the fault is apparently an extension of the Poso Creek rather than the Kern River fault.

TRAP FAULTS

The Mount Poso–Round Mountain fault (Fig. 2) is the trap fault of greatest length and displacement. It can be seen cutting the Temblor formations on the north bank of the Kern River, about 4 miles west of the Kern River scarp, and can be traced from topographic indications throughout a considerable part of its length.[16] Subsurface data show the displacements which account for traps for the Round Mountain, Mount Poso, and Mount Poso (Ring) pools. The fault plane dips to the east at an angle of about 60°; and, since the downthrow side is also to the east, a graben-like structure is produced with the boundary faults.

A number of small faults lie in the "graben," and these furnish traps for the Coffee Canyon, Dorsey, Pyramid Hill, Tampico, and north Mount Poso pools. Structurally these faults are the same as the Mount Poso–Round Mountain fault; they are normal with the downthrow side to the east.

AGE OF FAULTING

The boundary and trap faults were probably very nearly contemporaneous. Diepenbrock[17] states that in the Mount Poso area displacement of basal Kern River (Pliocene) formation along the Mount Poso fault is approximately the same as that of the Vedder (Miocene). Because the marine Etchegoin (Pliocene) formation pinches out west of the fault, it is impossible to differentiate the

[16] Recent completion of Shell's 56-x-11 Smith in 11-26-27 for production from Vedder indicates that the Mount Poso fault may continue several miles farther northwestward than shown on the accompanying diagram.

[17] "Mt. Poso Oil Field," *op. cit.*

Kern River and Chanac (Pliocene) formations. From the position and relationship of this pinchout, however, Diepenbrock's "basal Kern River" may be assumed to be early Pliocene, thus dating the faulting as post-early Pliocene. Although Fox[18] stated that sedimentary variations in the Round Mountain area were a result of intermittent fault movement throughout the Tertiary, additional subsurface data have failed to indicate diversities that could not be accounted for by strand-line fluctuations and source variations. It would seem, therefore, that the faults of the Kern River salient were formed in late Pliocene or Pleistocene times and thus correlate with the major uplift and faulting of the Sierra.

ORIGIN OF THE TRAP FAULTS

Over most of the area the sedimentary thickness is considerably less than 5,000 feet, and the sediments are poorly consolidated and relatively incompetent. Thus the sediments probably acted chiefly as a passive cover, merely reflecting basement movements. This, no doubt, accounts for the small amount of drag associated with the displacements.

Similar dying-out northwestward of the trap and boundary faults should be of genetic significance. Where surface traces of the Round Mountain and Kern River faults may be seen, the fault planes are found to intersect at depths of about 15,000 feet if the dips are extrapolated at 60°. Projections in other parts of the area also indicate intersections at similar depths. At such depths granodiorite is loaded far beneath its crushing strength, and it would seem that, because of the intersection, the origin of the trap faults may be directly dependent upon the boundary faults. The fact that maximum structural complexity exists in the southern half of the area, where the boundary faults overlap, is also indicative of this.

Mechanics of the formation of adjacent faults, whose planes intersect with depth, are somewhat speculative. The assumption made earlier—that the boundary faults were formed essentially as a result of shearing and tensional stresses initiated by a force couple—seems valid. However, formation of the subordinate trap faults requires consideration of several possibilities.

[18] *Op. cit.*

One hypothesis is to resolve compressive stresses along the boundary faults. These could have resulted from the expanding and doming effects of the Sierra, which was in process, or from adjustments caused by movements along the active San Andreas fault to the southwest. In either case the boundary faults antedate the trap faults; and tension, allowing a reversal in movement along the trap faults, would be necessary for the final adjustment of the blocks.

A more likely possibility is the development of conjugate shear planes as a result of a rotational force couple. Such a couple was undoubtedly involved in producing the uplift of the Sierra to the east and the subsidence of the valley floor to the west. With these vertical stresses initiating a rotational couple, conjugate shears would form at about the same time, the shear dipping to the west lying in the plane of the boundary faults, and that to the east in the plane of the trap faults. The boundary-fault shear should develop more completely and be the major fault because of its advantageous position with respect to subsequent stresses and frictional factors. Under such circumstances the trap faults are genetically dependent upon the boundary faults. The latter would be expected to form near the sedimentary-igneous contact, the hinge of the couple, and about which major displacements should continue.

Tension must be a dominant factor, for without it the amount of brecciation and distortion necessary for the known displacements on the trap faults is unreasonable. Assuming elasticity, tension, distortion, and rotation to be negligible, geometry of the system (see cross sections of Fig. 2) requires about a foot of brecciation, and subsequent shifting, at the intersection of the planes for each foot of displacement along the trap faults. However, tension, distortion, and elasticity would largely eliminate such brecciation, with tension allowing the "graben" block to be displaced vertically, much as a wedge, and elasticity distributing results of the necessary dilatation below the intersection of the fault planes, as well as allowing for compensating distortion.

The more acute the angle of intersection between the fault planes, the more difficult rotation becomes. Specifically, if the known dips of the fault planes are projected as linear curves, the angle of intersection approximates 60°. However, if these fault planes more nearly

approach a 45° dip with depth, as is probable, this angle approximates 90°, and rotation is easier. Moreover, curvature of the boundary-fault planes with depth would increase the friction resulting from additional movement and facilitate development of the trap faults.

A number of small faults, for the most part paralleling the regional grain, lie in the "graben." These were probably formed at the same time as, or perhaps shortly after, the outlining of the "graben" and, by making smaller blocks, aided movement and rotation.

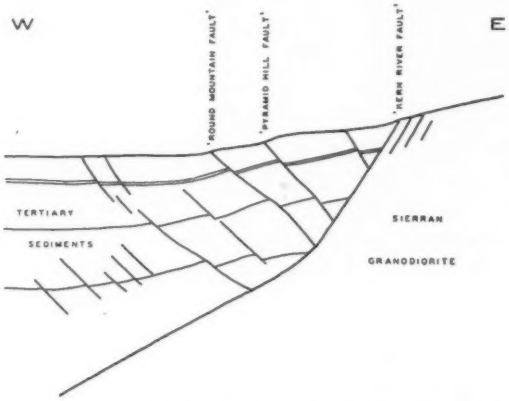

FIG. 3.—Modified drawing of one of H. Cloos's experiments specifically applied to this area.

Faults farther out in the valley at Premier, Kern Front, and Fruitvale are, no doubt, too distant to be directly associated with the boundary faults. These trap faults are of very small displacement and may result from adjustments necessitated by sedimentary inwash or Coast Range diastrophism.

Hans Cloos,[19] in his experimental work, has produced faults with the same structural relationships as those of the boundary and trap faults. Figure 3, which is a drawing of part of one of his experi-

[19] "Zur experimental Tecktonik," *Die Naturwiss.* Vol. XIX (1931), p. 242. "Kunstliche Gebirge," *Natur und Museum, Senckenbergische Naturwiss. Gesellsch.* (Frankfurt a. M.), Vol. V (1929), p. 227, and Vol. VI (1930), p. 259.

ments,[20] has been modified so as to be specifically applicable to this area. A comparison of this drawing and the cross sections of Figure 2 shows almost perfect agreement. In a duplication of Cloos's experiments, which consisted essentially in subjecting viscous mud to tension, incipient conjugate sets of shears were found to form almost simultaneously and, with further development, were controlled by adjustments along the major (boundary-fault) shear. Curvature of the fault plane and consequent drag were dominant factors in development of displacements in the fractures corresponding to the trap faults, and also in rotation and distortion of the "graben" blocks. Additional support for the conjugate-shear theory is found in the correspondence of the dips of both the experimental and actual faults with those expected for shears in a brittle substance such as granodiorite.

In addition to the aforementioned duplication of Cloos's experiments, repeated attempts were made to obtain such a fault system by different methods. Models were constructed in which fault-plane dips were 45°, 60°, 75°, and 90° and in which compression or tension could be applied at will. Scaled mediums varied from wet sharp sand through molding sand to various clays. It is thought that the generally poor results may be attributed to the extremely delicate balance necessary between tension caused by vertical movements and compression needed to keep the fault surfaces in contact. In Cloos's experiments, in which only horizontal tensile stresses were applied, this difficulty was obviated.

Developing of such faulting should be confined largely to rather homogenous competent masses, such as the Sierran. In a sedimentary series of considerable variation, stresses might be dissipated in adjustments along bedding planes and within more incompetent beds, rather than along the incipient shears.

Although such *subordinate conjugate faults* have not been commonly mentioned in the literature, infacing scarps have been described along the Wasatch Front and on the east side of the Alabama Hills in Owens Valley. As subordinate conjugate faults are always relatively small, and because they are located directly opposite and

[20] Cloos, "Zur experimental . . . . ," *op. cit.*, p. 243, Fig. 2.

adjacent to major faults, erosion and detrital filling would rapidly obscure their presence. L. F. Noble[21] has described a similar system associated with the West Kaibab fault in Muav-Flint canyon which serves as the boundary fault limiting the Unkar wedge on the northeast, in the Grand Canyon area. From subsurface data, the Pechelbronn oil district (Alsace), in the Rhine graben, has been found to parallel closely the faulting of the Kern River salient,[22] assuming the Black Forest boundary fault to be normal. In fact, Cloos specifically applied his experimental work, from which Figure 3 is modified, to Pechelbronn.

It would seem that more detailed work, particularly subsurface, may add other areas which have had a structural development similar to the Kern River salient. Subordinate conjugate faulting, which has produced graben-like structures, may not be, after all, so uncommon.

ACKNOWLEDGMENTS.—The writer is indebted to R. W. Clark and R. H. Miller, of the Western Gulf Oil Company, for aid in assembling material and for suggestions throughout the year-and-a-half spent in the area; and to Professors C. W. Merriam, O. D. von Engeln, and, particularly, C. M. Nevin, all of Cornell University, for assistance in the development of the paper.

[21] "The Shinumo Quadrangle," *U.S. Geol. Surv. Bull. 549* (1914), pp. 75–78.

[22] J. O. Haas and C. R. Hoffman, "Temperature Gradient in Pechelbronn Oil-bearing Region, Lower Alsace," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XIII (1929), pp. 1257–73.

UNUSUAL OLIVINE IN BASALT NEAR AUCKLAND NEW ZEALAND

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Within a radius of 12 miles from the center of the city of Auckland there is a cluster of about sixty basaltic cones or craters of Pleistocene to sub-Recent age, all composed of olivine-rich basalts in which there is relatively little variation in type but for rare basic limburgitic phases and the occurrence in one crater of ejected masses of a rock that approaches nepheline dolerite in character.

Determinations of the optic axial angles of the pyroxenes and olivines of representative examples of these Auckland basalts have been made with the universal stage by W. N. Benson and F. J. Turner,[1] but available analyses are rare.

A recently made thin section of an olivine basalt found as a chance boulder at Otahuhu, a suburb of Auckland, and not traceable to its parent-quarry, shows features in its olivine which are so unusual as to warrant record. This olivine is present as phenocrysts which rarely exceed 0.5 millimeter and often are below 0.15 millimeter in maximum dimension. A few of the olivine crystals are euhedral in outline, especially the smaller ones; and altogether they constitute about 13 per cent of the rock by volume. There is present a deep-brown glass (20–25 per cent) in which are enclosed lath-shaped crystals of plagioclase (25 per cent), a few small phenocrysts of zoned, brownish-green augite along with very numerous small grains of the same mineral (25–30 per cent), and tiny crystals and specks of magnetite (about 10 per cent). These percentages are by volume. During micrometric analysis it proved impracticable to estimate at all accurately pyroxene, magnetite, and glass, respectively, in the ma-

[1] "Comparative Composition-Variation Diagrams for the Cainozoic Igneous Rocks of New Zealand with Determinations of the Optic Axial Angles of the Pyroxenes and Olivines Contained Therein," *Trans. Roy. Soc. New Zealand*, Vol. LXIX (1939), pp. 56–72.

trix. Together they constitute 62 per cent by volume. The glass is patchy in its distribution; in some parts of the thin section it is relatively abundant, while in others it is almost excluded by closely appressed tiny grains of augite. This latter mineral is in very rare, larger, stumpy, zoned phenocrysts not more than 0.36 millimeter in maximum dimension and in more numerous smaller ones less than half this size, which grade in size into the small grains of the groundmass. In similar manner the plagioclase grades from laths as large as 1.25 millimeter×0.25 millimeter to tiny microlites.

The writer's thanks are due to Dr. F. J. Turner, of Otago University, Dunedin, who very kindly determined certain of the optical characters of the various minerals and obtained the following results: A large crystal of albite-twinned plagioclase has the composition An_{66}; the smaller laths may be less basic, for the writer obtained extinction angles in the zone perpendicular to (010) which suggest that the variety is An_{55}. The augite is a brownish-green zoned variety, probably titaniferous, which has unusually high values of the optic axial angle: in a small phenocryst $2V = 64°$ with an outer zone where $2V = 70°$, while a small grain in the groundmass gave $2V = 60°$. These values appear to suggest that most of the augite is diopsidic, with some soda content in the phenocrysts. For two larger phenocrysts of olivine Dr. Turner determined $2V = 90°$ and 86° (+), while a small crystal gave $2V = 90°$. These figures show that the mineral closely approaches forsterite in composition.

Although fresh and unaltered, the olivine shows an unusual feature which is believed to justify the publication of this note, namely, the ubiquitous separation of iron ore in minute irregular clots which often are aggregated into rodlike or stringlike bodies while the margins are bordered by a definite reaction zone of the same mineral. Figure 1 shows this separation in the manner typical of the larger phenocrysts; the smaller ones are less regular in outline and more closely beset by iron ore.

After examination of hundreds of thin sections of olivine basalts the writer can recall only two rocks in which the olivine showed similar characters. In one, it was demonstrated only by an occasional individual of the many olivine crystals present; in the other, which

represents scoria from the summit scoria cone of Rangitoto, a lava cone which forms an island near the entrance of the harbor at Auckland, separation of iron ore characterizes all the phenocrysts of olivine, but in manner less spectacular than in the Otahuhu rock figured. In other local basalts the olivine may contain sporadic small

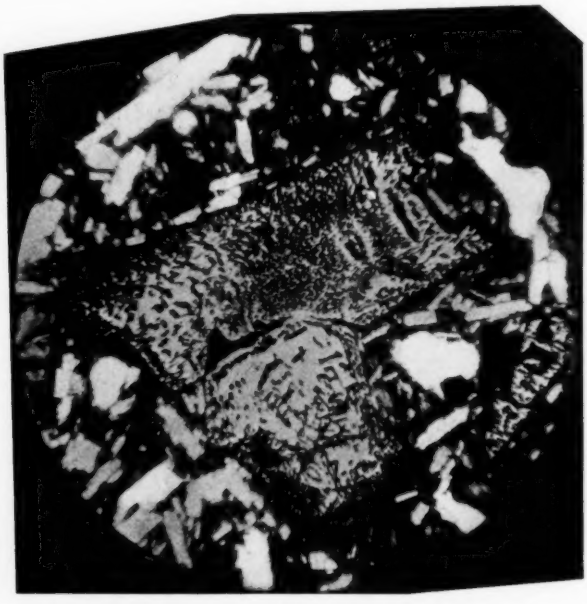

FIG. 1.—Fresh olivine (near forsterite) in olivine basalt from Otahuhu, Auckland, New Zealand, showing unusual separation of iron ore. ×65.

crystals of magnetite, but there is no reason for regarding them as other than normal inclusions.

In the case of the Rangitoto rock the few thin sections that the writer has made of more massive flows of basalt which form the main edifice of this cone (854 ft.) exhibit normal olivine;[2] the scoria with the abnormal olivine is the last eruptive product of the volcano. It appears, therefore, that the reaction—if it is such—that gave rise to

[2] *Ibid.*, p. 66, No. 7.

the iron ore occurred only at a very late stage in the eruptive history of the cone. On a single observation of optic-axial angle Benson and Turner[3] report enstatite from the groundmass of the Rangitoto basalt. It may be remarked, however, that small olivine bombs are abundant at a number of the Auckland volcanoes and that some include enstatite. In some instances the olivine and enstatite are in cataclastic granular condition, so that occasional enstatite amid general augite could well be explained by strewing of grains from such olivine bombs.

If the distribution and form of the iron ore in the crystals of olivine of Figure 1 are noted, it will be agreed that there is every suggestion that this iron ore has developed by reaction between early-formed crystals and liquid. If so, it is clear that ferriferous olivine has been converted to a variety near forsterite (as shown by Dr. Turner's determinations). If the facts at Rangitoto may be taken as a guide, this conversion took place at a very late stage in the volcanic cycle.

As pointed out by Dr. A. B. Edwards, of Melbourne, in a manuscript on the crystallization of basalts circulated in Australia and New Zealand in 1938, H. Kuno[4] and Edwards himself[5] have shown that the normal change in olivine during the progress of crystallization appears to be toward enrichment in FeO and impoverishment in MgO. The Rangitoto and Otahuhu rocks demonstrate, however, that such direction of change is not universal.

In conclusion, the writer would draw attention again to the unusual separation of iron ore that characterizes the olivine described; he must leave any deductions therefrom that may have a bearing upon crystallization in magmatic liquids to be drawn by others better qualified than he to discuss such matters.

[3] *Ibid.*

[4] "On the Crystallization of Pyroxenes from Rock Magmas with Special Reference to the Formation of Pigeonite," *Japanese Jour. Geol. and Geog.*, Vol. XIII (1936), pp. 141–50.

[5] "The Tertiary Volcanic Rocks of Central Victoria," *Quart. Jour. Geol. Soc. London*, Vol. XCIV (1938), pp. 243–320; "Tertiary Lavas from the Kerguelen Archipelago," *British Australia, New Zealand Antarct. Research Exped., 1929–1931*, Rept. Ser. A, Vol. II, Part V.

REVIEWS

Annual Reports of the Iowa Geological Survey, 1934–1939, with Accompanying Papers, Vol. XXVII. Des Moines, 1941. Pp. xiv+474.

This large volume includes the "Administrative Reports" for the years 1934–39 by A. C. Trowbridge; "Pleistocene Gravels of Iowa," George F. Kay; "Pollen Analysis of Interglacial Peats of Iowa," George H. Lane; "Geology of Adams County," Lyman W. Wood; "Mineral Production in Iowa for the Years 1933–38," H. Garland Hershey; and "A Summary of Mineral Production in Iowa, 1895–1938," by H. Garland Hershey.

More than two hundred pages are devoted to a masterly treatment of the Pleistocene gravels, interpretations of some of which have been basic in the gradual unfolding of the glacial history of our interior region. No state has contributed more than Iowa toward an understanding of this complex glacial story, and the various stages of advance toward present interpretations and concepts are so portrayed in this volume as to afford material of permanent value to the glacialist.

Further information on Aftonian and Sangamon interglacial conditions was obtained by determining the percentages of different plant genera from the fossil pollen in peat deposits. The Aftonian material indicates a coniferous vegetation gradually changing to grassland which prevailed for a long time. As the interglacial stage waned, grassland was succeeded by oak, particularly in the eastern part of the state, but oak dominance was short and conifers again became dominant.

Kay's thoroughgoing study of the Pleistocene gravels commences (p. 10) with a classification of the Pleistocene deposits of Iowa in which the Pleistocene is made a period, or system, which is subdivided into four epochs, or series, erected to cover groupings of the well-known glacial and interglacial stages. With this, however, the reviewer cannot agree, for reasons which he thinks should now be emphasized.

The shorter geologic periods have been about thirty million years in length, the longer ones perhaps twice that figure. Most periods comprise three or four epochs. Ten million years for an epoch, or the deposition of a series, is thus not above the average. The Pliocene was the last epoch of the Tertiary period. The new period, starting with the Pleistocene, we may call the "Quaternary," or perhaps get a better name for it. Its presumptive life is still some thirty million years into the future. How far has

the Pleistocene progressed? Its one million or two million years are far short of the ten million years of the average epoch. Today we are but fairly launched in the first epoch of the post-Tertiary period. The Pleistocene is but an epoch in that period, and we have as yet seen only a part of it.

Today the continents stand high in consequence of the marked diastrophism at the close of the Tertiary. Our present diversified climates are a correlative of this. As yet denudation really has only started to reduce the continents to extensive sea-level lowlands over which epeiric seas may spread widely as they so commonly have done in typical geologic periods of the past. But our present period still has time for that.

To regard the Pleistocene as a full-fledged period and to divide it into four epochs, or series (Grandian, Ottumwan, Centralian, and Eldoran), seem to be in striking disagreement with our classifications and interpretations of all preceding geologic history. Daly has urged thinking to scale, and here is a place to follow his precept. If the Pleistocene be properly only an epoch, or a part of an epoch, the so-called "epochs," Grandian, Ottumwan, Centralian, and Eldoran, disappear, and the Pleistocene series in the upper Mississippi Valley simply comprises the familiar glacial and interglacial stages—Nebraskan, Aftonian, Kansan, Yarmouth, Illinoian, Sangamon, Wisconsin, and Recent.

Is there any need for grouping each interglacial age with the preceding glacial age and giving to each combination a special name? The reviewer thinks not. Is the Yarmouth interglacial age so much more closely related to the preceding Kansan glacial age than to the succeeding Illinoian age that it should be allied with the Kansan and divorced from the Illinoian? Introduction of the terms "Grandian," "Ottumwan," "Centralian," and "Eldoran" has always seemed to the reviewer entirely unnecessary, putting an additional burden on students of the Pleistocene and serving to complicate rather than clarify the picture of Pleistocene history.

R. T. C.

Factors of Soil Formation: A System of Quantitative Pedology. By HANS JENNY. New York: McGraw-Hill Book Co., 1941. Pp. xii+281; figs. 125. $3.50.

In this volume Professor Jenny brings together the data of soil science in a manner markedly different from that of most treatises on the soil. Whereas most workers have attempted to co-ordinate the great mass of soil data by building all-embracing soil classifications, the present author

attacks the problem by attempting to assemble soil data into a comprehensive scheme based on numerical relationships. The approach in general is to assume that soil properties are functionally interrelated—that is, if one property changes, others also must change. If each measurable property of the soil is designated as s_i, this interrelation, expressed in the most general terms, is $F(s_1,s_2,s_3,\ldots)=0$. This general expression is broken down by choosing certain independent variables (climate, organisms, topography, parent-material, and time) as the soil-forming factors and by relating soil properties to them. Each s-property may thereupon be expressed as $s=f\,(cl, o,r,p,t)$, where the symbols in the parentheses refer to the soil-forming factors. In order to study the effect of each soil-forming factor on the individual properties, the other factors must be kept constant. The difficulty of satisfying this condition is pointed out by the author, but, despite its limitations, Professor Jenny marshals an imposing array of data which can be analyzed from the point of view developed in the book.

In his first chapter the author introduces basic definitions and develops the general functional relationship to be investigated. In succeeding chapters each soil-forming factor is examined in turn. Some of the more complex factors, such as climate, are resolved into subfactors, such as moisture and temperature. The treatment throughout is analytical, and as the text proceeds the reader begins to see an underlying pattern emerge from the mass of data on the soil. Numerous graphs showing the relation of soil properties to the soil-forming factors enhance the readability of the volume.

The reviewer is not a soil scientist, but as a geologist interested in the soil as an important part of the lithosphere he wishes to call the book to the attention especially of geologists who are interested in the continued quantification of the science. The book offers numerous suggestions for similar analytical attacks on more conventional geological problems. The concept that weathering processes follow definite stages of development to mature soils, which represent an approach to equilibrium conditions, is one that may be applied directly to geological studies.

Professor Jenny's book is not offered as a substitute for treatises which approach soil science from the classification viewpoint. As the author himself points out, both lines of attack supplement each other. The reviewer believes, however, that of the two attacks the present one offers the richer field for an ultimate understanding of the underlying principles which govern the formation and distribution of soils over the earth.

W. C. KRUMBEIN

Interpretation of Geologic Maps and Aerial Photos. By A. J. EARDLEY. Ann Arbor, Mich.: Lithoprinted and paper bound by Edwards Bros., 1941. (Distributed by the author, University of Michigan.) Pp. vii+99; figs. 40. $1.50.

This textbook is intended as a guide for laboratory work following introductory courses in geology, mineralogy, and petrology. The first thirty-eight pages are on ordinary geologic map work and are included to avoid the necessity of two manuals for such a course. This part contains thirteen figures, most of which are rather effective combinations of plan (map) and perspective block. It is divided into five sections: "Introduction," "Contacts" (including dip and thickness of beds), "Outcrop Patterns," "Structure Contours," and "Cross Sections."

Section 6 is a twelve-page general introduction to aerial photos. This contains plans for a simple mirror stereoscope. The reviewer suggests these be modified to include 50-cm.-focus positive meniscus lenses, one or two 8-watt fluorescent tubes along the sides, and first-surface mirrors. The faint double-images seen on either side of each object with ordinary mirrors cause undue retinal fatigue; this is especially noticeable when line drawings of stereopairs are being observed.

The next thirty-three pages deal with vertical aerial photos under the headings of contact tracing, drainage features, vegetation patterns, and deceptive markings. Many suggestive illustrations are given here. This section would be of much greater practical value had lists of the cheap aerial photos obtainable (in normal times) from the Department of Agriculture illustrating each type of feature discussed been included at the end of each heading or assembled at the rear. Many large mosaics are available from the government at a quarter of the cost of the listed commercial prices. Moreover, the county index maps obtainable with much smaller outlay are very effective in giving an even broader background picture than that from the ordinary mosaic. The reviewer would like to exchange such lists with others who have gone into this matter. After the correct discussion of shadows on page 50, it is surprising to find Figures 16, 23, 26, and 29 inverted.

The booklet concludes with a ten-page section on obliques, three brief exercises, and a short list of references. There is no index, but the table of contents is fairly detailed. Professor Eardley has done a real service in making this material generally available.

D. JEROME FISHER

Field Geology. By FREDERIC H. LAHEE. 4th ed. New York: McGraw-Hill Book Co., Inc., 1941. Pp. xxxii+853; figs. 599. $5.00.

The fourth edition of this well-known handbook differs from its predecessor, the third edition published in 1931, in no major way. The new edition retains the same format and style and essentially the same content arranged in the same manner.

The similarity in appearance and content, however, does not mean that no progress has been made in the decade that has elapsed since the earlier edition or that the author has not been aware of such progress. Many minor changes, mainly additions, have been made at various places in the text. The most important innovations are the addition of several pages on the primary internal structures of the igneous rocks, an appreciably enlarged section on airplane mapping with special reference to the use of the stereoscope, the addition of a section on electric surveying of drill-holes and the correlation of electric logs, and a considerable revision, with added illustrations, in the chapter on geophysical surveying. Sixty-one new illustrations have been added.

That this handbook has reached the fourth edition is ample proof of its usefulness to students and professional geologists alike. There is little doubt that this book will retain its deserved popularity. Any adverse comments on a fourth edition of a book so obviously well received may perhaps occasion surprise. Your reviewer, however, notes that the fourth edition does not include any methods of reconstruction of folds from observed strike and dip readings or the related problem of determination of thickness when the top and bottom surfaces of a bed have different dips (or even different strikes and dips), the two-tilt problem, or even the most useful, though older, two-level or contour method of W. S. T. Smith for the solution of fault problems.

F. J. PETTIJOHN

"Report of the Subcommittee on the Ecology of Marine Organisms." HARRY S. LADD, Chairman. Washington: National Research Council, 1941. Pp. 52. (Mimeographed.)

The first report of the Subcommittee on the Ecology of Marine Organisms, Committee on Geologic Research, has recently been issued by the National Research Council. It is a mimeographed bulletin of fifty-two pages, containing material presented last May at the annual meeting of the Division of Geology and Geography. Most of the data in the report were obtained by the subcommittee in a canvass of individuals and institutions interested in marine ecology as related to paleontology.

The report attempts to present a picture of current and recently completed activities together with an annotated bibliography of a number of recently published papers. The Division of Geology and Geography wishes to give the report a fairly wide distribution in the belief that it will be found useful by paleontologists. Copies have been sent to individuals and to libraries, particularly those in the geological departments of colleges and universities. A limited number of copies are on hand and will be mailed to those desiring them. Requests should be addressed to the Division of Geology and Geography, National Research Council, 2101 Constitution Avenue, Washington, D.C., accompanied by fifteen cents to cover cost of postage and handling.

WALTER H. BUCHER

Outlines of Structural Geology. By E. SHERBON HILLS. New York: Nordeman Publishing Co., 1940. Pp. ix+172; figs. 105; pls. 4. $2.25.

The aim of this author has been to present "a brief, yet reasonably complete and well-documented, summary of structural geology, with special reference to those aspects of the subject with which the field geologist should be acquainted." That, in fact, is just what he has done. Brevity has been accomplished by judicial selection and concise handling of material, yet in these relatively few pages the field of structural geology is well covered, from the opening chapter on "Non-diastrophic Structures" to the final one on "Petrofabric Analysis." The reader is conscious of getting the essentials with distinct economy of effort. The pervading dynamic point of view, by bringing out the significance of the descriptive details, makes them relatively easy to understand and remember. The book should prove practical either as the basis for a short course in structural geology or as a helpful reference book for a geologist or mining engineer to take with him in the field. Its small size (12mo.) favors the latter use.

The author, who is lecturer in geology in the University of Melbourne, expresses regret "that, owing to the geographical remoteness of Australia from the western countries, personal communication with all the authors whose figures have been reproduced . . . . has not been possible." The reader may make the mental note that, considering the many different sections of the globe represented, this would not have been overly easy, no matter where the author was located. The footnote references are unusually numerous and thoroughly cosmopolitan in scope. The many recent dates among them give a fresh flavor to the text.

R. T. C.

Fluorescent Light and Its Applications. By H. C. DAKE and JACK DE MENT. Brooklyn: Chemical Publishing Co., Inc., 1941. Pp. xiii+256; illus. 25. $3.00.

This small book covers all types of luminescence with particular emphasis on the practical side. It should be of especial interest to the amateur and the beginning student. After a historical introduction and an elementary treatment of radiation and visual perception, types and theory of luminescence are described. Much of the chapter on theory does not properly belong under this heading; in fact, the book is repetitious and not well organized. The next five chapters deal with technique and mineral fluorescence. A final chapter covers applications in other fields. The twenty-page bibliography is well-nigh valueless, as the arrangement under the thirteen headings is utterly haphazard. Thus names cited in the text are found (if present) in the bibliography only after a time-consuming search. There are author and subject indexes.

Repeated cases of bad English and bad science and many typographical errors were noted. Many of the equations presented contain unexplained symbols. In spite of these serious criticisms, there is much of interest in the book, and it should be of real value to the groups named.

D. J. F.

The Upper Ordovician Fauna of Frobisher Bay, Baffin Land. By SHARAT KUMAR ROY. "Field Museum of Natural History Geology Memoirs," Vol. II. Chicago: Field Museum of Natural History, 1941. Pp. 212; figs. 146.

The first portion of this work, written in an interesting travelogue style, deals with the narrative of the Rawson-Macmillan expedition to Labrador and Baffin Land. There are cited, in a preliminary fashion, some of the more prominent geologic features and processes noted by the author. Intensive work on most of these was prohibited by the exigencies of time.

The remainder of the paper includes a systematic description of the fossil fauna, a discussion of the several other collections from the area, and an inquiry into the age of "Silliman's Fossil Mount," the locality from which the collection was secured. The collection consists of 68 genera and 116 species and is by far the largest known from the region. Although the affinities of the fauna seem still to be questionable, Roy concludes that it is a recurring Mohawkian fauna of early Richmond age.

The Bibliography and Index are good, as are the figures in the systematic section.

ERNEST P. DU BOIS